Master Thesis in Geosciences

Observations on solifluction in Norway and Svalbard

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Cover photo: Solifluction lobes at Dovre, Norway.

Abstract

There have been many opinions on the definition of solifluction since the expression was first used by J. G. Andersson in 1906. The original definition was “slow flowing from higher to lower ground of masses of waste saturated with water. Later studies have revealed that there may be other factors to consider and, in addition, the soil does not necessarily have to be saturated. In this thesis, an overview of the existing research about solifluction has been made. Field work was carried out at Dovre and Svalbard.

Near Kapp Linné, Svalbard, two areas with widespread solifluction have been studied and described. At one study site, a 7 metre long trench was dug into the solifluction lobe, exposing a profile through the outermost part of the lobe. Fabric analysis showed that there is a preferred orientation of oblong stones, this is parallel to the direction of movement of the solifluction lobe.

At Dovre, similar studies were undertaken. By looking at aerial photographs, the distribution of solifluction in the field area was mapped. Fabric analysis done in one of the lobes showed the same results as in Svalbard: the preferred orientation for the oblong stones is with the longest axis in the same direction as the flow lines.

Temperature data from Fokstua meteorological station at Dovre has been compared with temperatures from two loggers set out at the field site. The loggers measured temperatures every hour during almost one year (2004-2005), situated at the tree line and on the ground near a solifluction lobe.

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Finally I would like to thank family and friends for all their support during my study.

Content

1 Introduction

Solifluction is a slow, downslope movement of soil taking place in the active layer. It happens in places where the ground is seasonal or perennially frozen, and involves both flow and creep processes (Benedict 1976).

Solifluction is a common phenomenon in periglacial regions, and is relatively easy to recognize in the landscape. A lot of research has been done to understand more about the processes, and one of the objectives of this study was to make an overview of the existing research. In order to obtain some visual information, field work was carried out in Svalbard and Dovre. The two field sites have different climate and ground conditions, it was therefore interesting to do field work in both places. Svalbard is situated in the arctic and mostly underlain permafrost. Dovre is in the inland of Norway, and there is discussion as to whether there has ever been permafrost here.

Due to the limited time for this thesis, no velocity rate investigations were undertaken. Instead, a profile of a solifluction lobe was closer examined in both field sites. Aerial photographs were used to get an overview of the distribution of solifluction in the areas.

All photos are taken by the author, unless mentioned in the figure text.

2 Theory

2.1 Definition

Solifluction is one of the most widespread movements of soil in periglacial regions. The expression was first used by J. G. Andersson to describe the “slow flowing from higher to lower ground of masses of waste saturated with water (this may come from snow-melting or rain)” that he observed at the Falkland Islands. He proposed the name solifluction which is derived from Latin *solum,* “soil”, and *fluere,* “to flow” (Andersson, 1906). The definition of solifluction has been changed over the years, and the scientists have had different opinions of the terminology. According to Matsuoka (2001), later studies have revealed that slow mass movements in cold regions include several processes, and do not necessarily have to be saturated. Other terms have been suggested, for example congelifluction (Dylik, 1951) for permafrost conditions, and gelifluction (e.g., Washburn, 1967) for frozen ground in general. Baulig (1956) defined gelifluction as solifluction associated with frozen ground. This included both permafrost and seasonally frozen ground. In modern usage, solifluction means slow mass movements associated with freeze-thaw action (Ballantyne and Harris, 1994; French, 1996). French (1976, 1996) suggests that solifluction is the total effect of frost creep and gelifluction, and this definition is in conventional use today. In the present thesis, this definition is adopted. Both frost creep and gelifluction derive from the effects of frost heave, caused by ice segregation within the frozen soil (Berthling, 2001).

To avoid the confusion using the terms gelifluction and solifluction, French (1996) points out the importance of whether the ground is seasonally (one-sided) or perennially (two-sided) frozen. Under one-sided freezing, solifluction consists of only two components: frost creep and gelifluction. The frost creep occurs usually in the autumn or in association with repeated diurnal frost throughout the year. Gelifluction occurs in the spring when the seasonally frozen ground thaws from the surface downwards.

The rate of solifluction is low, seldom more than a metre per year. Rate of movement is normally a few centimetres per year, but rates of 150-350 mm/yr have been recorded (Williams, 1959; Rapp, 1962; Holdgate, Allen and Chambers, 1967). Reasons for the variations in the flow rate are due to differences in relief, stone content, grain size, vegetation and snow cover (Veit et al., 1995). The soil discharge is normally between 15 and 114 $cm^3/cm/yr$ (Ballantyne, pers. comm., 2005). Benedict (1970) estimated the mean discharge in his study to be 59 $cm^3/cm/yr$. Other examples of estimated mean discharges are 26 $cm^3/cm/yr$ (Harris, 1981) and 50 $cm^3/cm/yr$ (Rapp, 1960).

Solifluction influences the denudation of mountains much less than rapid processes (such as landslides and mudflows) and geochemical transfers. Still, solifluction has a great effect on the evolution of mountain landscapes because it is so widespread. Solifluction is strongly dependent on climatic conditions, and can therefore be used as an indicator of the climate that has affected the slope in the past (Matsuoka, 2001).

Solifluction is favoured by sandy to silty soils having low liquid limits and plasticity indices (Harris, 1989). The exact timing of the movements in solifluction has never been established. The soil advances wavelike, in lobes, over the vegetation cover. It is still uncertain why it behaves like this, and exactly how it moves. When the soil moves over the vegetation cover it leaves a buried organic layer. This layer can provide an estimate of the rate and extent of the movement. Such layers may cover many metres and can represent vegetation overrun during thousands of years. The fronts of such waves, terraces or lobes often contain an accumulation of boulders and stones. The reason for this is still not known (Williams and Smith, 1989).

The best condition for solifluction to occur is in the presence of an impermeable layer, such as frozen ground, which limits the downward movement of water through the soil. Most effective is the process where melting ice layers weaken the soil and provide a source of excess water that reduces internal friction and cohesion (Williams, 1957b, 1959). The frequency of the freeze-thaw action and the depth and thickness of ice lenses are the major controls on the spatial variation in solifluction processes (Matsuoka, 2001).

Rates and processes of solifluction are dependent on climate, hydrology, topography and the geology in the area. Prediction of landscape evolution in periglacial mountains requires quantitative relationships between the rate of solifluction and these variables (Lewkowicz, 1988; Kirkby, 1995). Field measurements have been done for this purpose, and large amounts of data have been collected from polar hillslopes (e.g., Washburn, 1967) to tropical high mountains (e.g., Francou and Bertran, 1997).

French (1996) has collected published data from several researchers, and made a table that shows rates of solifluction movements (table 2.1.1). As seen in this table, the rates vary between 0.02 and 12.0 cm/yr from place to place. They also vary greatly within each locality. The highest rates are found in Spitsbergen, the lowest in Canada.

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Table 2.1.1 Some recorded rates of solifluction movement (French, 1996)

Locality	Reference	Gradient (degrees)	Rate (cm/year)
Spitsbergen	Jahn (1960)	3–4	1.0–3.0
Spitsbergen	Jahn (1961)	7–15	5.0–12.0
Kärkevagge, Sweden	Rapp (1960a)	15	4.0
Tarna area, Sweden	Rudberg (1962)	5	0.9–1.8
Norra Storfjell, Sweden	Rudberg (1964)	5	0.9–3.8
French Alps	Pissart (1964)		1.0
East Greenland	Washburn (1967)		
		'Wet' sites	3.7
		'Dry' sites	0.9
Colorado Rockies	Benedict (1970)		0.4–4.3
Okstindan, Norway	Harris (1972)	5–17	1.0–6.0
Ruby Range, YT, Canada	Price (1973)	14–18	0.6–3.5
Sachs Harbour, Banks Island, NWT, Canada	French (1974a)	3	1.5–2.0
Garry Island, NWT, Canada	Mackay (1981)	1–7	0.4–1.0
Swiss Alps	Gamper (1983)		0.02–0.1
Eastern Banks Island, NWT, Canada	Egginton and French (1985)	<10	0.6
Svalbard	Akerman (1993)	2–25	
		Stripes	1.4–2.0
		Steps	0.9–2.2
		Lobes	3.3
		Sheets	4.3

2.2 Classification

Matsuoka (2001) classifies solifluction into frost creep, needle ice creep, gelifluction and plug-like flow. Frost creep is the downslope movement of soil particles originating from frost heaving normal to the slope followed by nearly vertical thaw consolidation (Washburn, 1979). It can be subdivided into diurnal frost creep and annual frost creep in accordance with the period for the freeze-thaw cycle. The difference in frost duration affects the depth of ice lenses developing during frost heaving, and thus controls the depth of movement (fig. 2.2.1B and C). Diurnal frost creep only affects the uppermost centimetres of soil (Matsuoka, 1998a), while annual frost creep usually occurs down to a few decimetres depth (e.g., Smith, 1988).

Needle ice creep is a special kind of diurnal frost creep. It occurs when surface debris (one to a few grains thick) is lifted by ice needles and falls on thawing (fig. 2.2.1A). The growth of ice needles reflects nocturnal cooling to just below 0°C, when the frost plane will go no deeper than the uppermost centimetre of soil. Such a thermal condition mainly occurs in early winter when the subsurface soil is still warm enough to prevent nocturnal frost penetration (Matsuoka, 2001).

Gelifluction is mainly associated with seasonal thawing (fig. 2.2.1C) (Matsuoka, 2001). Suitable conditions for this are places where the downward percolation of water through the soil is limited because of the frozen soil. Gelifluction mainly operates in the thawing-season. The best localities for this to happen are places where the moisture is very low, for example under late-laying or perennial snow. The movement is usually limited to the uppermost 50 cm of the active layer (French, 1996). Identification of the gelifluction component in the total soil movement is often difficult, since it may be confused with annual frost creep and retrograde components (Matsuoka, 2001).

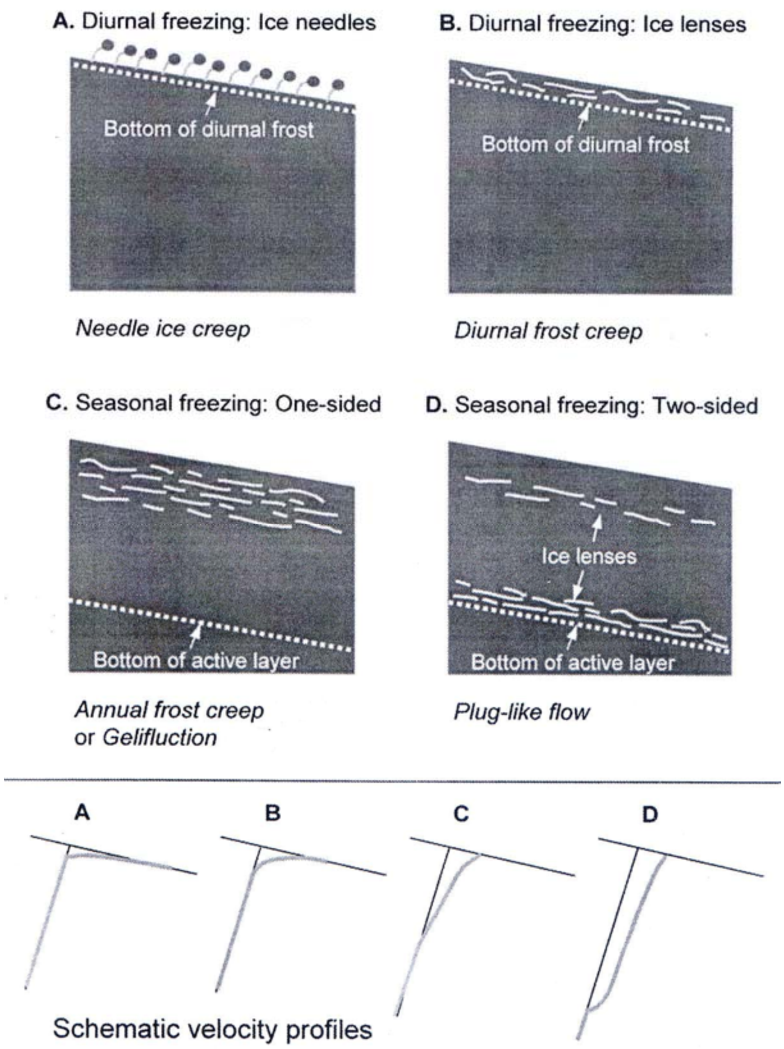


Fig. 2.2.1. Types of frost heave and solifluction (Moderated from Matsuoka, 2001).

The last type of solifluction is plug-like flow (fig. 2.2.1D), this type only exists in cold permafrost slopes and should be distinguished from gelifluction. Under two-sided freezing, solifluction includes not only frost creep and gelifluction, but also a movement which occurs in late summer. At this time the thawed active layer is capable of sliding across the slip plane provided by the ice-rich zone at the top of permafrost. This is termed “plug-like” flow (Mackay, 1981).

In warm permafrost areas and areas without permafrost one-sided freezing (downwards) normally dominates. No matter how deep the frost plane penetrates, dehydration caused by ice segregation in the upper soil hinders ice lens growth at depth. In contrast, two-sided (both downward and upward) freezing can occur in the presence of cold permafrost, because a large thermal gradient permits the frost to penetrate upwards from the top of permafrost (Lewkowicz, 1988). The upward freezing may produce numerous ice lenses near the base of the frozen active layer, which on thawing induce movement of the whole active layer as a plug (fig. 2.2.1D) (Matsuoka, 2001).

According to Washburn (1967), solifluction consists of gelifluction and frost creep acting together. The relation of the two features is shown in figure 2.2.2. In addition, there is a retrograde movement that is not fully understood. This is an upslope component of movement that probably is related to cohesive forces in the drying soil, or to capillary pressures (Washburn, 1967). The movement is the upslope component of movement caused by nonvertical settling of the soil during thaw (Benedict, 1976).

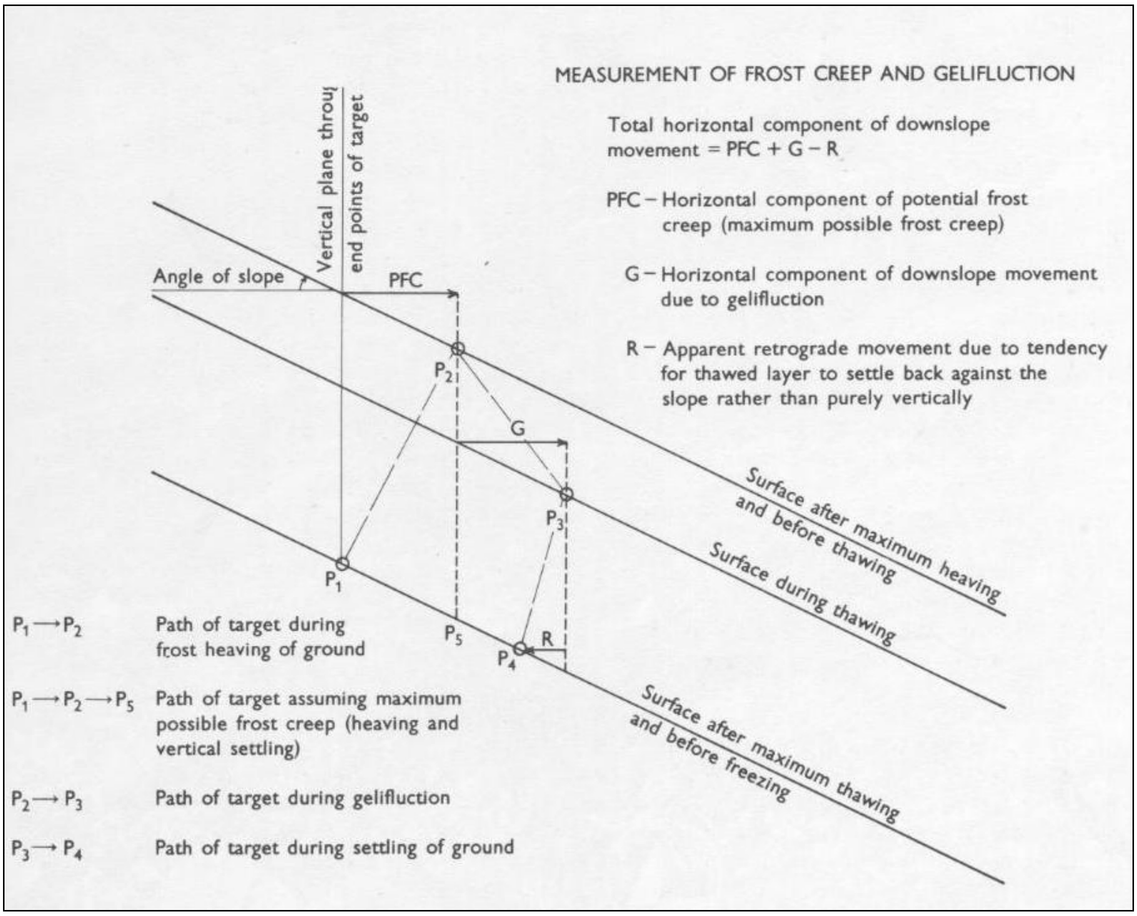


Figure 2.2.2. Relation of gelifluction to frost creep and retrograde movement with reference to a vertical plane through the end points of a line at right angles to movement (Washburn, 1967).

2.3 Landforms

Solifluction is visible in many forms, such as lobes, terraces (or steps), sheets, stripes and hummocks. Figure 2.3.1 shows some of these forms. The solifluction lobe is a relatively well known phenomenon, and is found in most periglacial environments. Usually the lobe is a result of both gelifluction and frost creep (French, 1996). Washburn (1967) noted a dependence of moisture, with higher displacements in wet sites than dry. In wetter areas, gelifluction is more dominant; in dryer areas creep is the dominant factor. According to French (1996), the lobes occurs in smooth terrain, often the angles can be as low as 1-3°. The lobes are capable of transporting large erratic boulders, referred to as ploughing blocks. These blocks are often resting at or near the permafrost table, and sometimes they leave a shallow trough upslope indicating the path of their movement. Solifluction sheets are probably best developed in alpine areas where there is little vegetation; this allows the sheet to develop uniformly. In the lowlands and in the woods more local sheet movements can be seen. Solifluction terraces will develop where the movement rate is uniform in a large area, while lobes will develop where the movement is channelized. There are many intergradations in shape (Benedict, 1976).

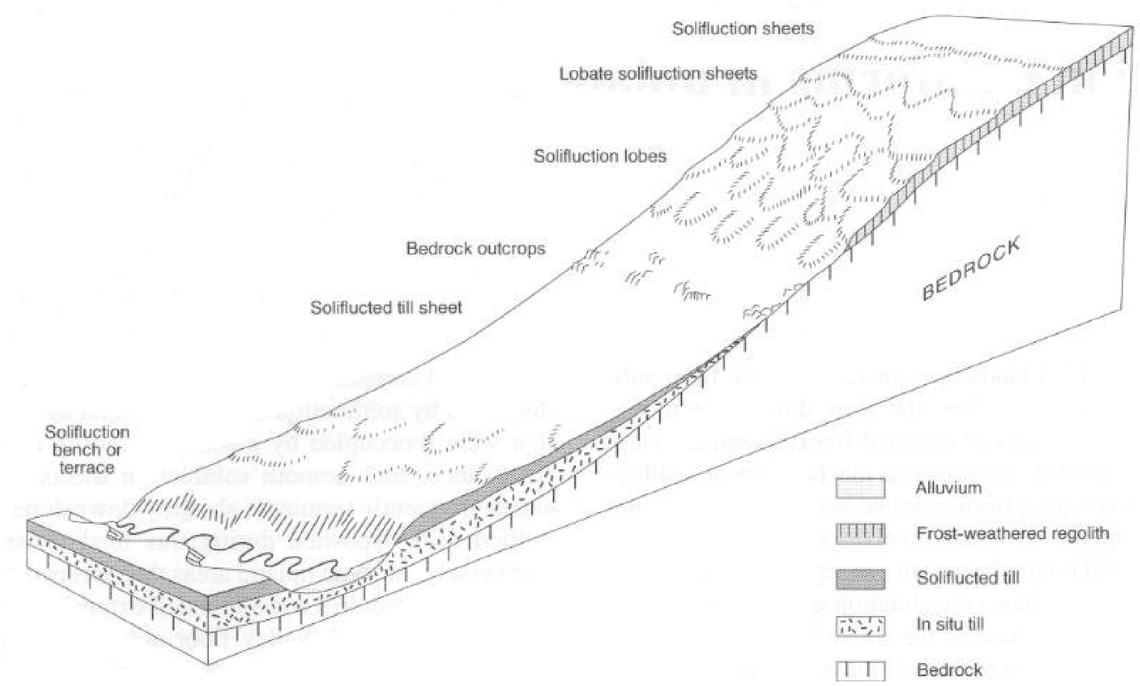


Fig. 2.3.1. A schematic catena of solifluction phenomena (Ballantyne and Harris, 1994).

A catena of solifluction phenomena like the one showed in figure 2.3.1 is visible in many upland areas. Smooth solifluction sheets occupy the gentle upper slopes, and often completely bury the underlying bedrock with their continuous debris mantle. When the gradient increases beyond about 5º, such solifluction sheets often terminate downslope in regular steps or risers, which run continuously along the slope for tens or hundreds of metres (Ballantyne and Harris, 1994). With further increases in gradient, the planform of such risers becomes increasingly crenulate or lobate in response to local variations in the rate of movement of debris. They will therefore form lobate solifluction sheets, and many of the features referred to as solifluction lobes represent the lobate extensions of such sheets. Farther downslope, lobate sheets tend to override each other, isolating individual lobes from their neighbours. On the steep central portions of mountain slopes, solifluction lobes tend to die out downslope and are replaced by a smooth, unbroken sheet of soliflucted debris (Ballantyne and Harris, 1994).

Lobes and terraces can be either turf-banked (fig. 2.3.2) or stone-banked (fig. 2.3.3), most common are the turf-banked. They develop where strong vertical frost sorting is absent, and where rates of soil movement decrease downslope, causing the soil to thicken. The groundwater regime will change after the initial thickening has occurred. Then the restraining effect of a rim of relatively dry soil at the front of the lobe or terrace encourages continued growth. Vegetation can assist the growth process, because it has a restraining effect upon movement (Benedict, 1976). According to Ballantyne and Harris (1994), turf-banked terraces that reflect retardation of creeping debris by bands of vegetation have developed in response to the combined action of mass-movement and the effects of wind on vegetation cover. Turf-banked lobes and terraces are normally associated with a decrease in gradient such as occurs on concave lower slopes (Benedict, 1970).

Fig. 2.3.2. Turf-banked lobe, Dovre. This lobe is situated in the field area in the present paper. See section 3.1 for more details.

Stone-banked lobes and terraces are less widespread because they require specialized conditions for their formations. Like turf-banked lobes and terraces, they form in response to a downslope decrease in the velocity of moving soil. Sorting can develop as a by-product of frost creep, but in many cases appears to have been inherited from previous intervals of vertical frost sorting. Stone-banked lobes develop at the front of stone stripes and stone streams. Terraces develop at the downslope margins of moving blockfields, and lobate terraces where entire flights of stone stripes spread laterally and coalesce. The features can form only where coarse debris is moving more rapidly than fines (Benedict, 1970). Although data is sparse, studies indicate that stones can be expected to move faster than fines only where movement is dominated by frost creep (Benedict, 1976).

Fig. 2.3.3. Stone-banked lobe from the western side of Griegaksla, Svalbard.
The solifluction at this site is discussed closer in section 3.2.

Frost creep and gelifluction features occur on slopes with angles between 2° and 25-35°. Benedict (1976) explains their absence on steeper gradients due to a rapid drainage of moisture. The reason can also be an inability of the deposits to maintain their fronts when they are too steep.

Buried organic matter is found in almost all lobes and terraces. The layers occur when soil humus is overrun by the deposit during its downslope advance. Sometimes more than one layer is present because several lobes have overridden each other. Benedict (1976) points out the importance to remember that humus horizons can also be preserved in alpine mudflow and landslide deposits.

Gelifluction is dependent of depth, gradient of the slope, amount of liquid during the spring thaw and grain size of the soil (Harris, 1973). The vegetation cover is also a controlling factor for soil movement rates. Harris (1973) found that the movement at the surface was dependent on the vegetation cover. Gamper (1983) and Smith (1987) found that the vegetation prevented movement because the upper layer was held together by the roots. The fastest movement was found where the vegetation was only a thin moss cover. Others have stated that vegetation actually favours slope movements, as the vegetation helps retain water and retards run-off (Sigafoos and Hopkins, 1952; cf. Washburn, 1979).

2.4 Solifluction as a climate indicator

Frost creep and gelifluction commonly operate together. The importance of each process varies from site to site and from year to year, and has varied, on a much larger scale, during the climatic changes of late Pleistocene and recent time (Benedict, 1970). Due to this, solifluction features can give us information about the palaeoclimate. Even if we recognize frost creep and gelifluction features in the geological record, we can not be sure if there used to be permafrost at the location. However, a distinction must be made between activity and origin. Lobes and terraces are currently active in non-permafrost environments, but this not necessarily means that they are forming in these environments under present conditions (Benedict, 1976).

Frost creep and gelifluction are favoured by seasonal frost however permafrost is not a compulsory requirement. Most radiocarbon-dated lobes and terraces seem to have originated either during, or at the close of, intervals in which shallow permafrost was more widespread than today (Benedict, 1976). Between ca. 3000 and 2500 B.P, growing evidence show that lobes and terraces developed in many parts of the world. This period is believed to have been colder than the present, with glacier advances (Denton and Karlén, 1973).

While fresh and active, solifluction features are generally easy to recognize, unfortunately, after some time they can be difficult to distinguish from other mass-wasting processes (Benedict, 1976).

2.5 Previous field studies

Many experiments have been done in the field, one example is a study that Benedict (1970) did in the Colorado Front Range. During eight years, soil-movement processes in a hillside above the tree line were observed in order to describe the origins, rates and former rates of the moving soil.

Periodic measurements of downslope movement and frost during the 1965-66 annual freeze-thaw cycle suggest that solifluction is a more effective process than frost creep in the saturated axial areas of turf-banked lobes in wet sites. At the edges, frost creep is more effective. In this study, Benedict (1970) uses the term solifluction about all downslope displacement occurring during the spring thaw. Frost creep also occurred during this period, but was too shallow to affect the movement of wooden stakes used in measuring displacement.

Rates of displacement were strongly influenced by differences in moisture availability and slope gradient, but were relatively unaffected by differences in soil texture and temperature. Maximum rates of downslope movement measured in sorted stripes, turf-bank lobes and terraces, and stone-banked lobes and terraces at nine experimental sites on Niwot Range varied from 0.4 to 4.3 cm/year. Movement is currently confined to the upper 50 cm of soil; columns of small cement rods placed in vertical drillholes showed no downslope displacement below this depth after four years of burial (Benedict, 1970).

At one experimental site, potential frost creep exceeded theoretical values calculated from heave and slope measurements. In addition, the retrograde movement was larger than expected. At a site where frost creep is the dominant movement process, stone-banked lobes moved three times as fast as the finer-textured soil between them. At a site where shallow solifluction is important, stripes of coarse debris moved only half as rapidly as stripes of fine material (Benedict, 1970).

The most comprehensive field studies so far are those of Matsuoka (1994, 1996, 1998) and Matsuoka et al. (1997). In these studies, continuous measurements were done of both horizontal and vertical displacements, as well as soil temperatures and soil moisture at several depths. However, the gelifluction mechanisms could not be evaluated further because frost creep dominated the study sites.

2.6 Previous laboratory experiments

According to Harris et al. (1996), field studies have not yet been precise enough to allow a more detailed exploration of the mechanisms of periglacial mass movement. An alternative approach is the simulation of periglacial slope processes where a slope is constructed in the laboratory and subjected to a controlled thermal and hydraulic regime. Laboratory experimentation does not only give us greater precision and detail of monitoring. By reducing the annual cycle of freezing and thawing to a few weeks, it also allows rapid simulation of the equivalent of many years soil movement over a period of only a few months.

Higashi and Corte (1971) were among the first to recognize that the relationship between frost creep and gelifluction may be explored through controlled laboratory modelling. They undertook slope simulation experiments using a tilted box (67cm x 42 cm x 15 cm), filled with a 10 cm layer of very frost-susceptible silty clay soil. Water was supplied during freezing through wicks from the base. The air temperature above the slope model surface was controlled by a refrigeration coil and heating element. Soil temperatures were monitored, and soil surface displacements were measured manually at the end of each freeze-thaw cycle using markers. Markers were also buried at depth increments of 2 cm and their displacement measured after two or three freezing cycles (Higashi and Corte, 1971).

By relating the rate of surface frost heaving to the depth of frost penetration, frozen moisture contents were calculated. Experiments were performed using a constant pattern of temperature cycling with slope angles of 3°, 7.5° and 15°. Soil freezing was accompanied by the growth of needle ice that lifted surface markers and those at 2 cm depth. Ice needles were observed to bend downslope under their own weight, and cause downslope rotation and sliding of markers as they thawed. In addition, the release of meltwater into the thawing soil caused slow saturated flow (gelifluction). Through the experiment Higashi and Corte (1971) found out that the downslope displacements resulting from near-surface frost creep and gelifluction were proportional to the square of the slope, so that slope angle was identified as the dominant factor influencing surface movement rates (Higashi and Corte, 1971).

3 Field work

In order to gain additional information about solifluction, field work was done at Svalbard and Dovre. In both places, a trench was dug into a solifluction lobe, and fabric analysis was done. Visual studies were undertaken, in order to get an overview of the area. As a result of this, a geomorphological map was made.

3.1 Dovre

Due to the predominant solifluction features close to the main road, Dovre was chosen as a field site. The site had several advantages: It is easily accessible, it has a weather station nearby (approximately 7 km from the field site) and it is within reach from Oslo. The weather station is situated at Fokstua (at about 62º06’ N, 09º17’ E), 952 m a.s.l. This station was built in 1923, and a continuous meteorological record since 1957 has been used in this study.

3.1.1 The study site

The study site situated in Dovre municipality in Oppland county (fig. 3.1.1), is in the southern part of Norway (at about 62º09’ N, 09º27’ E). Located between Dombås and Hjerkinn, about 3 km south of the main road, a hillside is clearly visible with solifluction lobes. The area under closer examination is between two rivers. Field studies were carried out in autumn 2004, spring and autumn 2005.

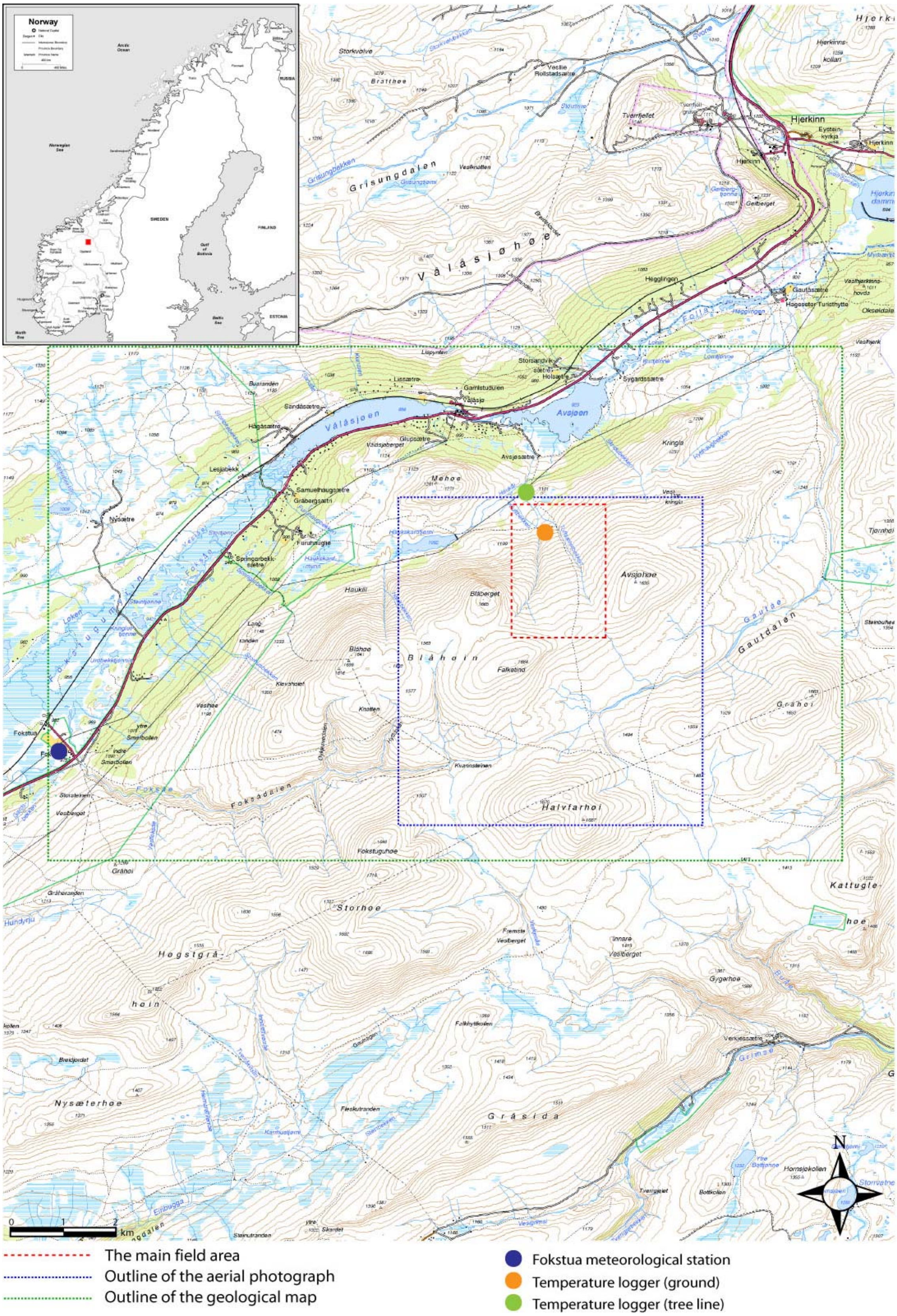


Fig. 3.1.1. Map of the field area.
(Location map, Norway: MapArt. Cartesia Software 1994, USA
Topographical map: Statens kartverk, 1992. Norge 1:50000.
Topografisk hovedkartserie-M711. Blad 1519 III Hjerkinn)

3.1.2 Collecting data

Three data loggers measuring the temperatures were set out in the beginning of October 2004 and retrieved in the beginning of September 2005. The loggers were measuring temperatures once an hour during this period. One logger was put in a tree at the tree line, approximately 1.5 metre above the ground. Another logger was laid at the ground at the side of one of the solifluction lobes in the area above the tree line (fig. 3.1.1). The logger was put at a site expected to be snow-covered first. The third logger was placed on top of the solifluction lobe, at the same lobe as where the second logger was situated. Unfortunately, the third logger malfunctioned, therefore no data could be obtained. In figure 3.1.1, the locations of the loggers are marked.

By comparing the temperatures from the data loggers with the data from the meteorological station, it is possible to say something about the meteorological record of the field site.

3.1.3 Topography

The study area is situated in a hillside facing towards NNW (fig. 3.1.2). As is common in Dovre, the landscape is very open, with smooth mountains. In this study, an area located above the tree line between two rivers was shown most attention. The altitude in this area is between 1050 and 1600 m a.s.l., the average slope angle ca. 10-15º, with small variations.

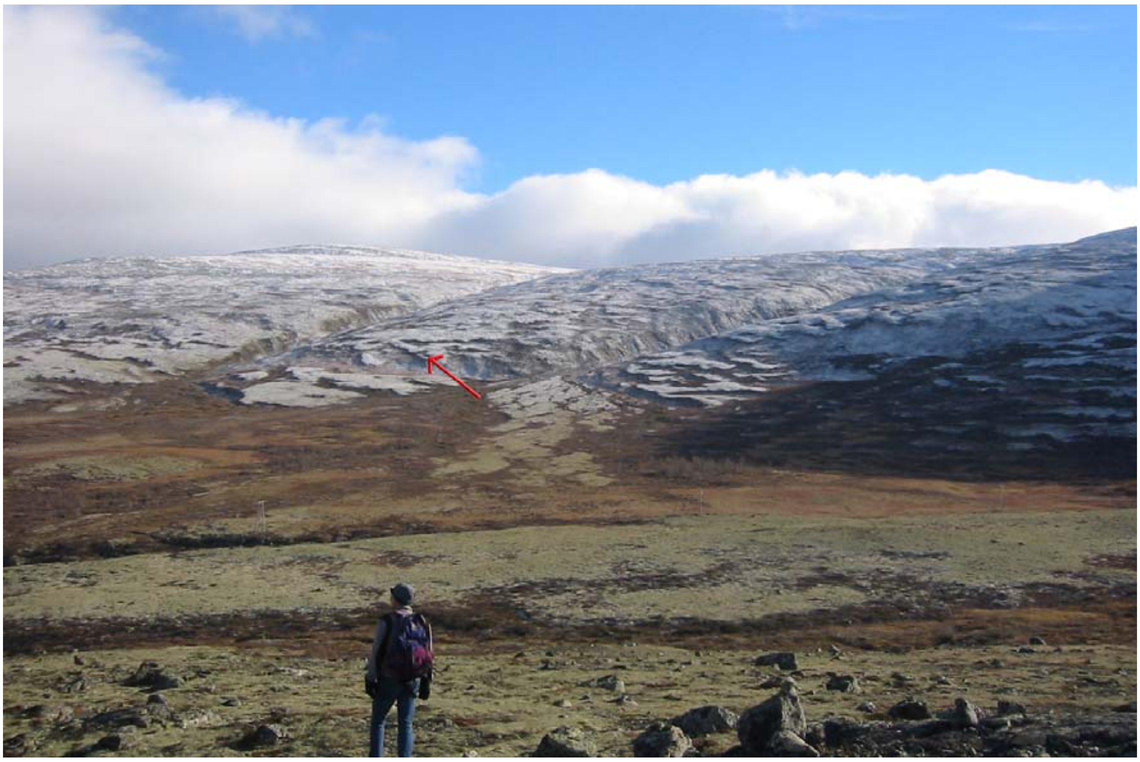


Fig. 3.1.2. View towards the study site. The picture was taken in October 2004, when the solifluction lobes are covered with a thin layer of snow. An arrow indicates the location of the ground temperature logger.

3.1.4 Present and Holocene climate

Present climate

The climate at Dovre is continental, with dry, cold winters and wet, warm summers. Average temperature in January-February is about -10ºC, in June-August about +10ºC. Fokstua meteorological station is situated at 952 m a.s.l, about 7 kilometres from the field site. The mean annual air temperature (MAAT) in 1961-1990 is -0.7ºC, and the annual precipitation is 450 mm. During the period 1957-2004 the average wind speed is 4.3 m/s, coming from 170º (Data from the Norwegian meteorological institute).

Figure 3.1.3 compares the temperatures from the meteorological station Fokstua with the temperatures from the data loggers at the field site. The logger at the tree line is situated at 1040 m a.s.l.; the logger at the solifluction lobe is situated at 1098 m a.s.l. As seen in the graph, the temperature at the tree line is 2-3 degrees lower than the meteorological station in winter time. Between March and September the temperatures are approximately the same. The temperature at the solifluction lobe follows the temperature at the tree line when the ground is free of snow. Covered with snow, the temperature remains relatively stable at about -1ºC to -3ºC. In the same period, the temperature at the tree line varies between -23ºC and 7ºC.

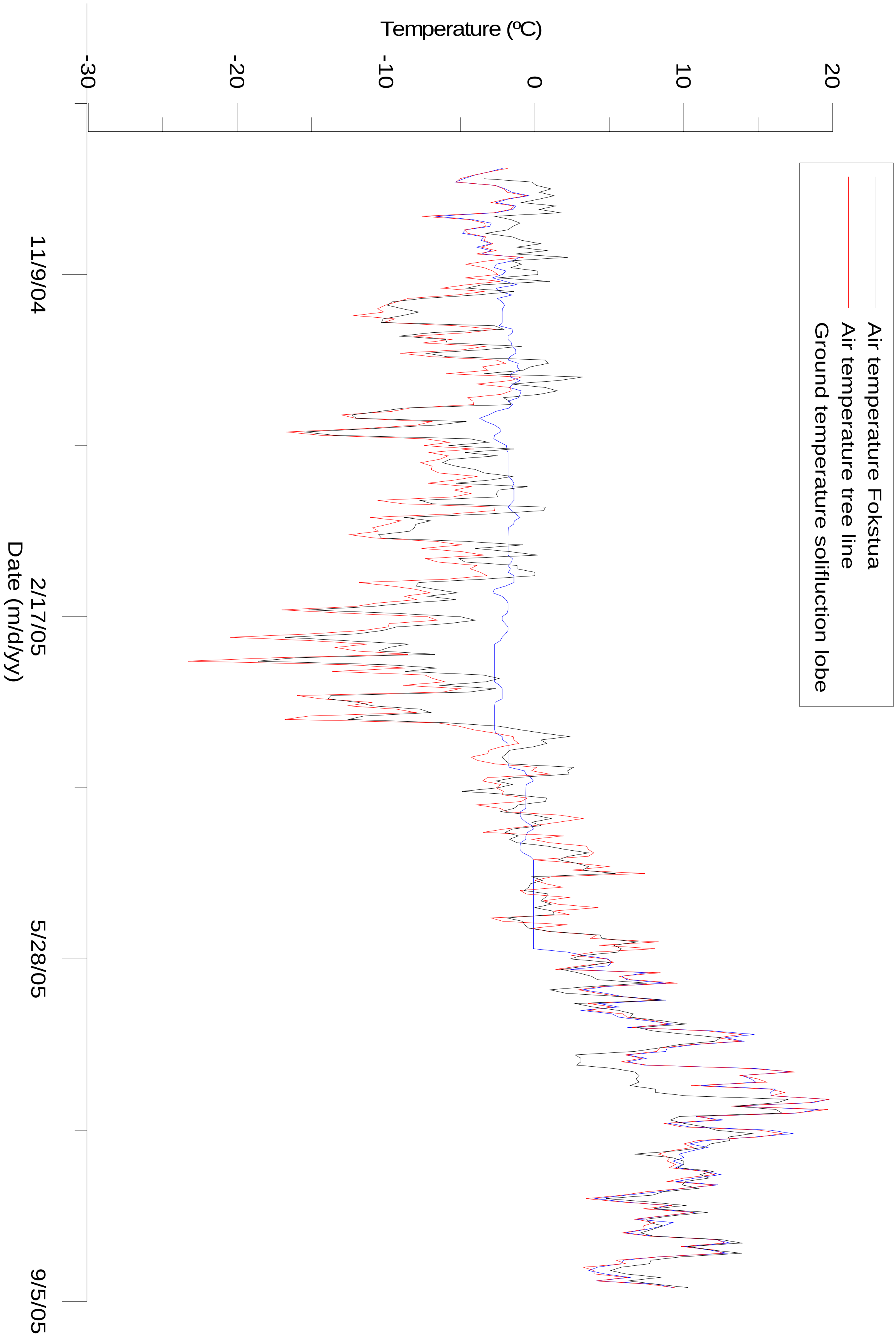


Fig. 3.1.3. Temperature variations at the field site and the meteorological station between Oct 8 2004 and Sept 1 2005

Holocene climate

The Holocene is considered to be the period after the last glacial maximum until now, e.g. 11,500 years B.P. until present. There were three global periods with glacier expansion in this period: in 5800-4900 B.P., 3300-2400 B.P. and in 1500-1900 A.D. The last one is called the Little Ice Age. In Swedish Lapland there was also an earlier glacial expansion at about 8000 years B.P.

Between ca. 950 and 1200 A.D. there is evidence that the climate was much warmer than now. This period is referred to as the Little Optimum (Frakes, 1979).

The early Holocene (appr. 4500-12,000 B.P.) is believed to be warmer than the last 4500 years. However, some of the warmth may primarily represent seasonal warmth rather than year-round warmth (Frakes, 1979). 6000 years B.P., mean July temperatures across Europe were warmer than present. Areas of high elevation in central Europe may have been warmer by as much as 5ºC. Mid-Norway had an increase of 0-2ºC (Huntley and Prentice, 1988). In the late Holocene (appr. 3500-4500 years B.P.), the climate was cooler and dryer. This period is sometimes known as the Neoglaciation.

In order to find out more about the possibility of permafrost in the field area, a graph was made (fig. 3.1.4). The graph shows in which altitude one will find a MAAT of 0ºC and -2ºC. By using the temperatures measured at Fokstua meteorological station between 1957 and 2004, and a lapse rate of -0.6ºC/100m, the following formula was made:

Altitude with 0ºC: $952 + \left(\frac{100 * T_F}{0.6} \right)$

Altitude with -2ºC: $952+\left(\frac{100*T_F+200}{0.6}\right)$

where T_F is the temperature at Fokstua, situated at 952 m a.s.l. These calculations were done for each year, giving the following graph.

The graph showing the altitude with MAAT of 0ºC (fig. 3.1.4) gives an indication of where to find permafrost. Permafrost usually occurs at places with MAAT below 0ºC, and the graph shows that an altitude of at least 650 m has to be reached to find possible permafrost between 1957 and 2004. The altitude varies between 650 and 1200 m during the period, and the tendency is that the altitude is rising. It is possible to state that there is permafrost in the field area as it is situated between 1100-1600 m a.s.l.

The graph showing the altitude with MAAT of -2ºC (fig. 3.1.4) gives an indication of where to find permafrost in the Holocene. As mentioned, the climate was probably 2 degrees warmer than present. The graph will therefore indicate on which altitude the MAAT was 0ºC in the Holocene. As seen in the graph, the altitude must be more than 1000 m before this limit is reached, and the altitude varies between 1000 and 1550 m in the calculated period.

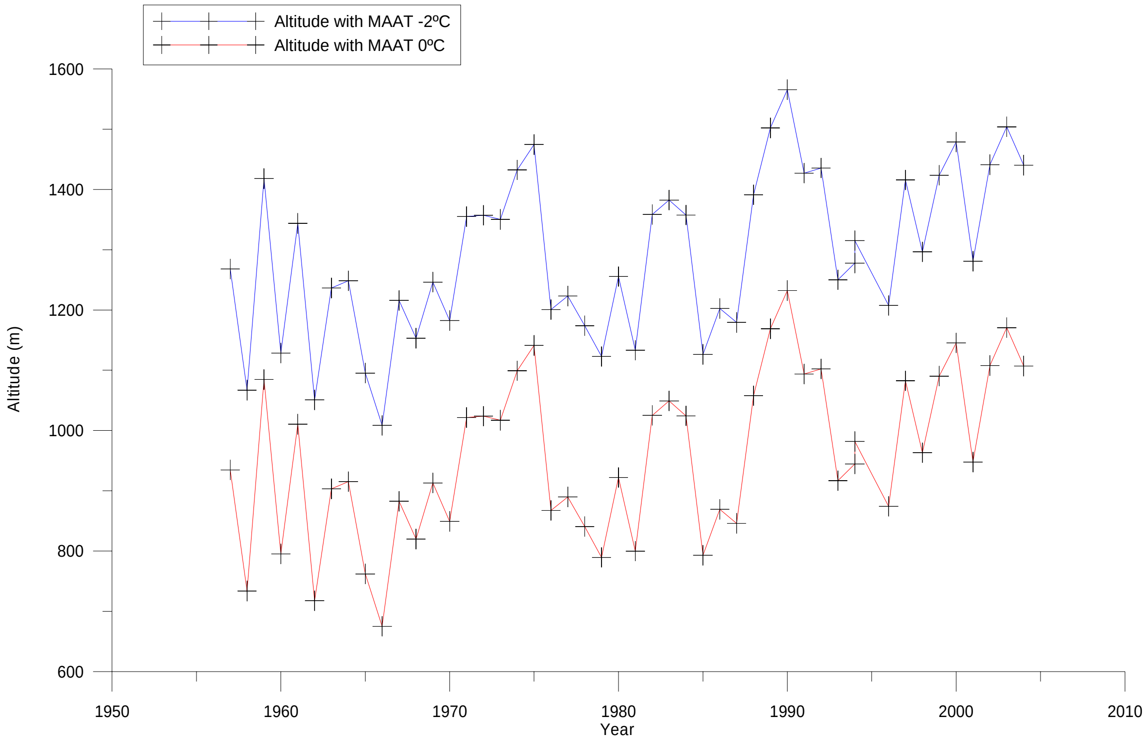


Fig. 3.1.4. Graph showing the altitude of a MAAT of 0ºC and -2ºC during 1957-2004, based on temperatures from Fokstua meteorological station. The red 0ºC-line gives an indication of where to find possible permafrost at the present time. The blue -2ºC-line shows where to find possible permafrost in the Holocene.

3.1.5 Solifluction in the study area

The studied hillside has a view towards NNW. The highest point in this area with solifluction is ca. 1580 m a.s.l., the lowest point about 1150 m a.s.l. An aerial photograph of the area is shown in figure 3.1.6. In this area, solifluction lobes are dominating the mountain hill. The lobes have different dimensions, measured between 4 and 44 metres long, and 7 and 34 metres wide. They have a gradient of 10-20º on top of the lobe, which steepens until about 22-32º towards the end. Some of them were longer than they were wide, others were wider than they were long. The thickness of the lobes was normally 1-3 metres. Figure 3.1.5 shows a typical solifluction lobe in the area. In this lobe, fabric measurements were carried out, the results are discussed in chapter 3.1.9.

Fig. 3.1.5 Example of a typical solifluction lobe in the area. This lobe is 35 m long, 28 m wide and about 3 m thick at the thickest point. The gradient is 30º at the end, and 12º at the top.

Previous research of solifluction has been done in Dovre. Near Snøhetta, approximately 22 kilometres from the field site, instrumental methods showed a solifluction displacement of up to 25 cm during a single spring (Holtedahl, 1960). Near Reinheim, Williams (1957a) investigated some well-developed solifluction terraces on slopes of 5º-20º, probably situated on a large outwash fan. He also examined some terraces in Trollheimen. At this site, the terraces occur on morainic and talus deposits at the foot of the south-east scarp of a mountain. Due to a 13-metre long humus layer found under one of the solifluction terraces, Williams (1957a) concluded that the front must have advanced over at least that distance in post-glacial time. In this area that means perhaps 6000 years, which gives a movement rate of at least 3 mm/yr.

Fig. 3.1.6. Aerial photograph of the field area. The solifluction lobes are marked with red.
(Photo: Fjellanger Widerøe, 1963. Series 1431 J18)

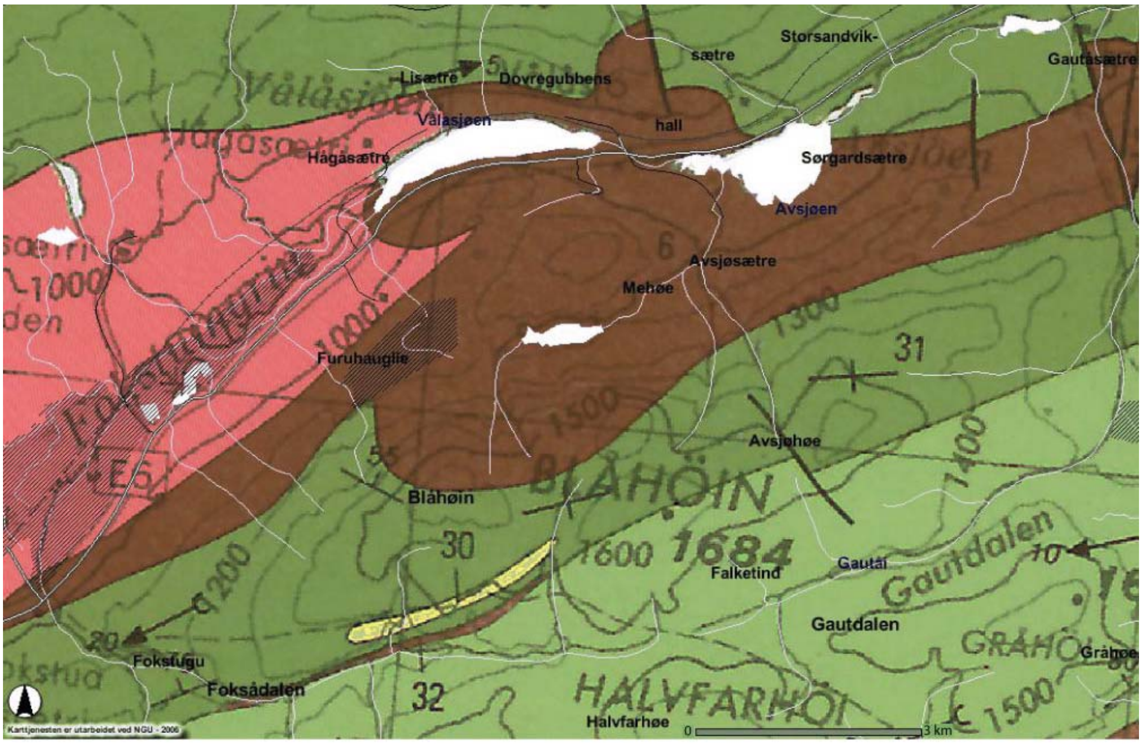
3.1.6 Vegetation

The study site is situated above the tree line, which is at about 1040 m a.s.l. in this area. Birch dominates the forest below this elevation. The ground is covered with moss, lichen and some small bushes. In the lowermost part of the field, at about 1150 m a.s.l., reindeer moss is the dominating species. On the lowermost lobes there are several species of lichen, moss, salics vier and crowberry. Outside the lobes, blueberry bushes and dwarf birch dominate. Further uphill, blueberry and moss are less frequent and lichen dominates.

3.1.7 Geology

The bedrock at Dovre belongs to two different bedrock-complexes. In the western part, old (more than 600 million years) Precambrian bedrock is present. This part is dominated by acid hard rocks like gneiss and feldspar-rich sandstone. The eastern part belongs to Trondheimsfeltet, a part of the Caledonian mountain range. This mountain range was formed 400-600 million years ago (www.uib.no/bot/qeprg/dovre.htm).

The place where the fieldwork was done consists of sediments from Ordovician and Silurian (diorite, gabbro and metagabbro) in the lowest part, between 925 and 1300 m a.s.l. Between 1300 and 1665 m a.s.l., the ground consists of rocks from Proterozoic to Cambrian, grey phyllite, biotite phyllite and micaschist. See figure 3.1.7 for more details.



Grey phyllite, biotite phyllite and -shale

Quartz micaschist

Diorite, gabbro, metagabbro

Amphibolite

Trondhjemite, granite and monzonite

Conglomerate with quartzite

Fig. 3.1.7. Geological map of the area
(Digital version of: Nilsen, O. and Wolff, F.C., 1989. Geological map of Norway, Rørås og Sveg 1:250000. NGU. URL: http://www.ngu.no File: NGU248470 (date: 25.01.06))

3.1.8 Snow distribution and depth

The weather station is approximately 7 km from the study site, and gives a good indication of the conditions in the field. In winter 2004/2005, a thin snow cover developed in the middle of October 2004. This cover mostly disappeared due to warmer days in late October. Only small patches prevailed in the depressions. After this period, new snow fell and stayed during the whole winter. The thickness of the snow cover was up to 75 cm during the winter, and covering 100% of the ground most of the time. The last snow at the weather station disappeared in the end of May 2005, probably a bit later at the field site.

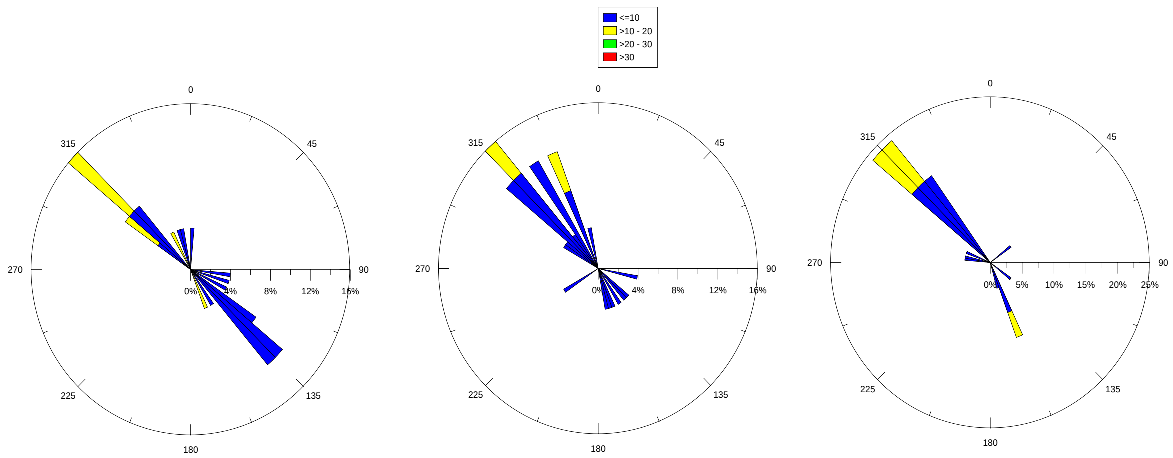
The period of snow cover determined from the data logger situated by the solifluction lobe is from the middle of November until May 26. This is the period where the temperature was stable.

Field studies in the end of May 2005 gave more information of how the snow melts around the solifluction lobes. By this time, the lobe surfaces were snow free, however snow could still be found on the ground surrounding the lobes to a depth of 1.5 m (fig. 3.1.8). Most places, the snow layer was 60-70 cm thick.

Fig. 3.1.8. Solifluction lobe at Dovre in the middle of November 2005. As seen in the picture, the solifluction lobe is free of snow, whereas the area around is snow covered.

3.1.9 Fabric analysis

At one of the lobes situated at about 1200 m a.s.l., a 90 cm deep hole was dug. It was dug 10 m from the end of the lobe, parallel to the flow direction. At three levels, 25 oblong stones were measured. Their dimensions were noted, as well as their dip and preferred orientation. Based on the orientation of the length axis, a rose diagram was made for each level. These diagrams are shown in fig. 3.1.9.



upper 10 cm middle (40-50 cm depth) lower (80-90 cm depth)

Fig. 3.1.9. Rose diagram showing the orientation and dip of the longest axis of the oblong stones in the trench. Blue colour indicates that the dip is equal to or less than 10º.
Stones dipping 10-20º are marked with yellow.

As seen in the rose diagram, the stones have a preferred orientation. They tend to have their longest axis towards either ca. 315º or 135º, most of them towards ca. 315º. This corresponds well to the orientation of the lobe itself, which is pointing towards approximately 330º. A majority of the stones, more than 90%, have a dip of 10º or less, the rest have a dip of 10-15º.

At the lowermost level, 36% of the stones are dipping towards 312-325º. The trend is quite clear at this level: preferred orientation is ca. 320º.

In the middle level, 25% of the stones have a preferred orientation of 312-320º. 51% of the stones are dipping towards 312-340º. Opposite this side, 26% have a dip towards 135-160º. Also at this level, the trend is towards ca. 320º.

In the uppermost part of the profile, 52% of the stones are orientated towards 100-160º. 48% have a dip towards 314-360º. No stones are dipping towards the other directions, so a trend is obvious.

3.1.10 Discussion

Solifluction is a phenomenon that is favoured by permafrost, but not required. Whether or not permafrost is present at Dovre, is still unclear. With a MAAT of around 0ºC at the meteorological station, the conditions do not favour permafrost. The MAAT should be some degrees lower for optimal permafrost conditions. However, the meteorological record is limited, so former climate conditions might have made permafrost possible. According to figure 3.1.4, the field site is situated well above the 0ºC limit both in the last decades as well as in the calculation for Holocene.

About 5 kilometres west of the field site, palsas are found. Palsas are peaty permafrost mounds that are normally 1-7 metres high, and less than 100 metres in diameter. They consist of alternating layers of segregated ice and peat or mineral soil material. According to French (1996), palsas are one of the few reliable surface indicators of permafrost in the discontinuous zone. The palsas near the field site indicate that there is possibly at least discontinuous permafrost at Dovre.

Fabric analysis shows that the oblong stones have a preferred orientation that corresponds well to the flow direction of the flow. This indicates that the flow must have been going in the same direction for some time. The trend is more distinct in the lowermost part of the profile, with a majority pointing towards NW. However, in the uppermost level the stones are still pointing towards either NW or SE, with a difference of 180º.

3.2 Svalbard

3.2.1 Study objectives

The objective of this study was to obtain knowledge on solifluction in an arctic environment. The study site was at Kapp Linné (78º04' N, 13º38' E) on the west coast of Svalbard (see map in fig. 3.2.1), which has a wide range of interesting glacial and periglacial features. The fieldwork took place during 10 days in late July 2005.

3.2.2 The study site

<u>Location</u>
The field site was Kapp Linné, on the west coast of Spitsbergen. Kapp Linné is situated at 78º04'N, 13º38' E, at the southern shore of Isfjorden. At Kapp Linné there is a meteorological station at Isfjord Radio. This station is situated 50 metres from the sea, at 7 m a.s.l.

Solifluction phenomena are widespread along the foot of the Griegaksla mountain, located about 10 km SE of Isfjord Radio. The maximum altitude is 778 m a.s.l. Two study sites on either side (west and east) were chosen for more detailed investigations. A location map is shown in figure 3.2.1, with the study area marked with a triangle.

<u>Climate</u>

The climate at Kapp Linné is arctic, but with maritime influence. According to French (1976), the high arctic climate in polar latitudes has an extremely weak diurnal pattern, but a strong seasonal pattern. This is highlighted by the small daily but large annual temperature range. The maritime influence reduces the extreme temperatures, and also brings more precipitation than generally is experienced at similar latitudes (Humlum et al. 2003). Since Kapp Linné is situated on the border between a maritime arctic and a more continental arctic climate, the local variations are great. The mean annual air temperature at Isfjorden Radio for the period 1912-1975 is -4.8ºC; the mean annual precipitation is around 400 mm for the same period (Åkerman, 2005).

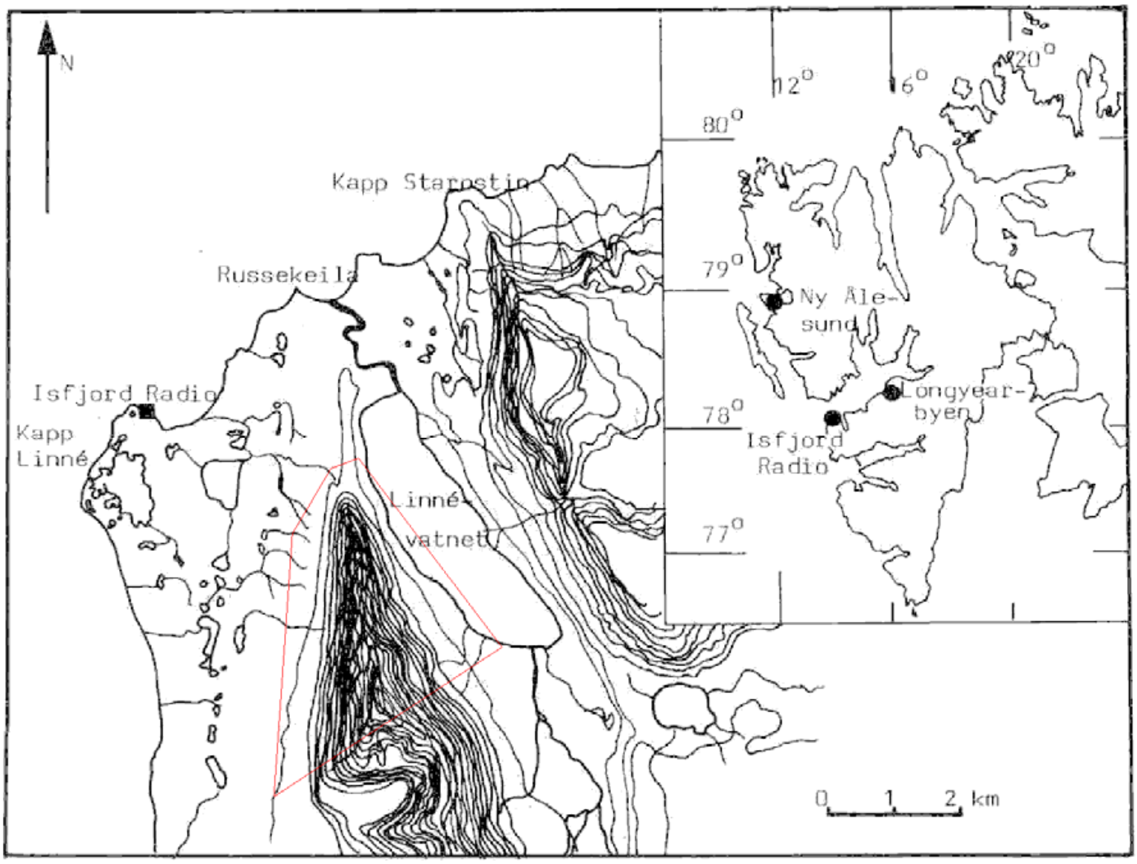


Fig. 3.2.1. Location map. The study area by Griegaksla marked with a triangle (Modified from Åkerman, 1980).

Geology

Fig. 3.2.2 shows a geological map of the area. As seen on the map, the eastern part of Griegaksla is dominated by sandstone. The western part of the mountain consists of phyllites (Hecla Hoek) from the Precambrian-Ordovician that strikes more or less parallel to the coastline (Åkerman, 1980, 1996).

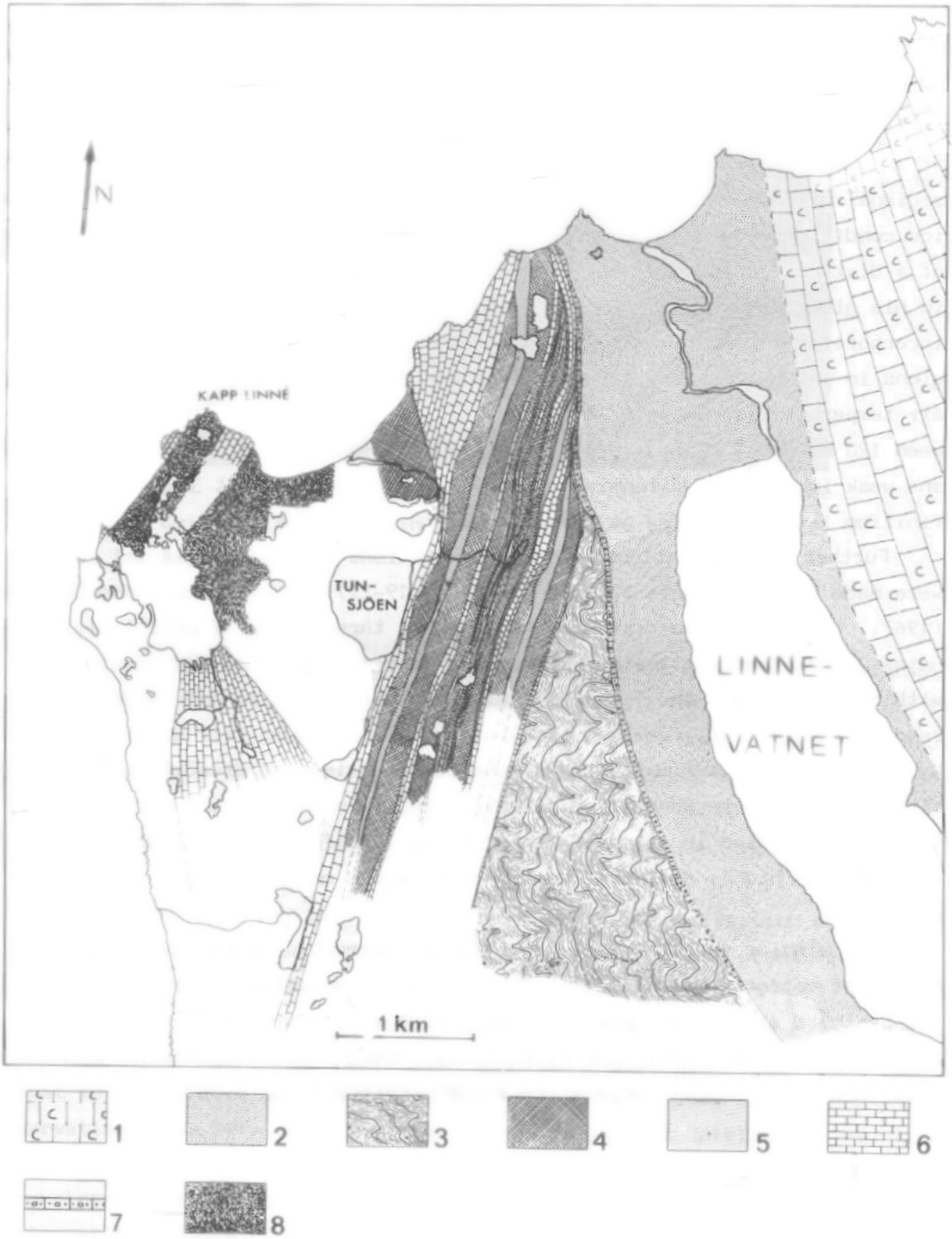


Fig. 3.2.2. Geological map of the investigation area (Åkerman, 1980).
1. Cyathophyllum limestone(middle Carboniferous), 2. Sandstones (lower Carbonifireous),
3. Phyllites (Hecla Hoek), 4. Schists (Hecla Hoek), 5. Quartzite (Hecla Hoek),
6. Dolomitic limestone (Hecla Hoek), 7. Konglomerate (Hecla Hoek) and 8. Tillites (Hecla Hoek).

Deglaciation history and coast line development during Holocene

Based on the geology of the sea bed east of Svalbard, and more than a 100 m Holocene rebound on Kong Karls Land, Landvik et al. (1987) suggest that an ice sheet existed in the northern Barents Sea during the Late Weichselian. In the sea west of Spitsbergen, moraine lobes are found in the submarine extensions of Kongsfjorden and Isfjorden. This is the best evidence for the coverage by glacier ice of most of the archipelago (Ohta 1982; cf. Landvik et al. 1987).

Permafrost in the area

With a mean annual air temperature of -4.8ºC, the Kapp Linné area is suitable for permafrost. The distribution and depth of the permafrost in the area is unclear, because the major parts of the boreholes are restricted to the central part of Spitsbergen and the inner part of the fiords. However, all observations indicate that the Svalbard region lies well within the zone of continuous permafrost, and according to Åkerman (1980, 1996) all observations have shown that the entire Kapp Linné area is underlain with permafrost. Even close to the coastline, permafrost is present. Åkerman (1996) has observed the temperature in two shallow drillholes, and estimated the depth of the permafrost to ca. 80 metres at a place situated 1500 metres from the coastline. The temperature of the permafrost at the level of zero annual amplitude was -4.9ºC.

Recorded depth of permafrost in the Svalbard region is between 75 and 450 metres (Liestøl 1977; cf. Åkerman 1980). As far as is known, permafrost covers the whole of Svalbard. The only exception is below the majority of the glaciers (Åkerman 1980, 1996; Liestøl 1977).

Previous investigations

The area around Griegaksla has been investigated by Åkerman during several years. He started investigating the area in 1972, and has done work there until recently. Åkerman noted that the solifluction sheets need a deep and well-established thawed active layer to be fully active, and are the largest features that need this.

According to Åkerman (2005) the first movements of solifluction lobes, sorted steps and sorted stripes at Kapp Linné occur in mid-late June. To be fully active, the depth of thaw should be more than 30 cm. During the very warm and very cool summers the correlations between the rate of movement and the air (and ground) temperature are good. The variability during “normal” summers is high, but this only tells us that the air temperature is one of several controlling factors. In July it seems like the number of rainy days are important, but not the precipitation amounts.

Åkerman (2005) investigated solifluction sheets and lobes during 1972-2002 on either side of Griegaksla. He distinguished between lobes and sheets, and measured the downslope surface displacement. For the period 1972- 1992, the lobes moved with a speed of 34 mm/yr and the sheets had a speed of 68 mm/yr. For the period 1992-2002, the speed of the lobes had increased to 81 mm/yr but the speed of the sheets remained constant at 68 mm/yr in the same period.

Matsuoka and Hirakawa (2000) have done some studies of soil movement at Kapp Linné. They installed two probes near the front of a lobe situated at 30 m a.s.l. Excavation of the probes showed that the maximum depth of movement was 40-60 cm during the two years of investigation. The velocity profile showed downslope convexity above 20-30 cm depth and concavity below. During the two years of field studies, the surface movement reached approximately 5 cm, which gives a velocity of ca. 2.5 cm/yr. The active layer thickness was estimated to be about 150 ± 20 cm, estimated from isotherms.

South of Kapp Linné, near Bellsund, Repelewska-Pękalowa and Pękala (1993) did studies of the active layer in 1986-1992. Among other things, they measured the amount of solifluction movement on slopes exposed to the north, south, east, and west. The inclination was about 10º, the slope was covered with Quaternary gravelly and sandy sediments. Thickness of the active layer was ca. 87 cm in the slopes facing north- and westwards, and 111-116 cm in the slopes facing towards south and east. The rate varied between 0 and 40 cm/yr, the largest variation was on the south-facing slope. At this slope, the mean rate was 6.5 cm/yr, which was the highest mean rate registered in this study. The lowest mean rate was found in the north-facing slope, with a mean movement of 4.0 cm/yr. In the slope facing towards east, the solifluction rate was 4.6 cm/yr, and in the slope facing westwards, the rate was 5.3 cm/yr.

3.2.3 Field work

Different methods were used in order to get an overview of the solifluction in the area around Griegaksla. Two field sites representing different types of solifluction were inspected visually, and a geomorphological map was made. In one of the solifluction lobes on the western side of the mountain, a trench was dug. To obtain information on the character of sediment deformation, fabric analysis was carried out. In order to estimate the age of the lobe, a Schmidt hammer was applied.

<u>Mapping the areas</u>
With the help of aerial photographs and field studies, a geomorphological map of the field area was made. The map of the eastern side shows the extension of the solifluction, talus and beach deposits, and can be seen in figure 3.2.3. There were no satisfying aerial photographs of the western side of the mountain; the picture in figure 3.2.3 has a shadow in this area. Another aerial photo of the wanted area exists, but the resolution was too poor for this purpose. The best solution in this case was to use an overview picture of the solifluction lobe. The outline of the lobe has been marked, and can be seen in figure 3.2.7. On the picture, sorted stripes are visible. By looking at them it is possible to see the direction of the flow, these flowlines are indicated in figure 3.2.4.

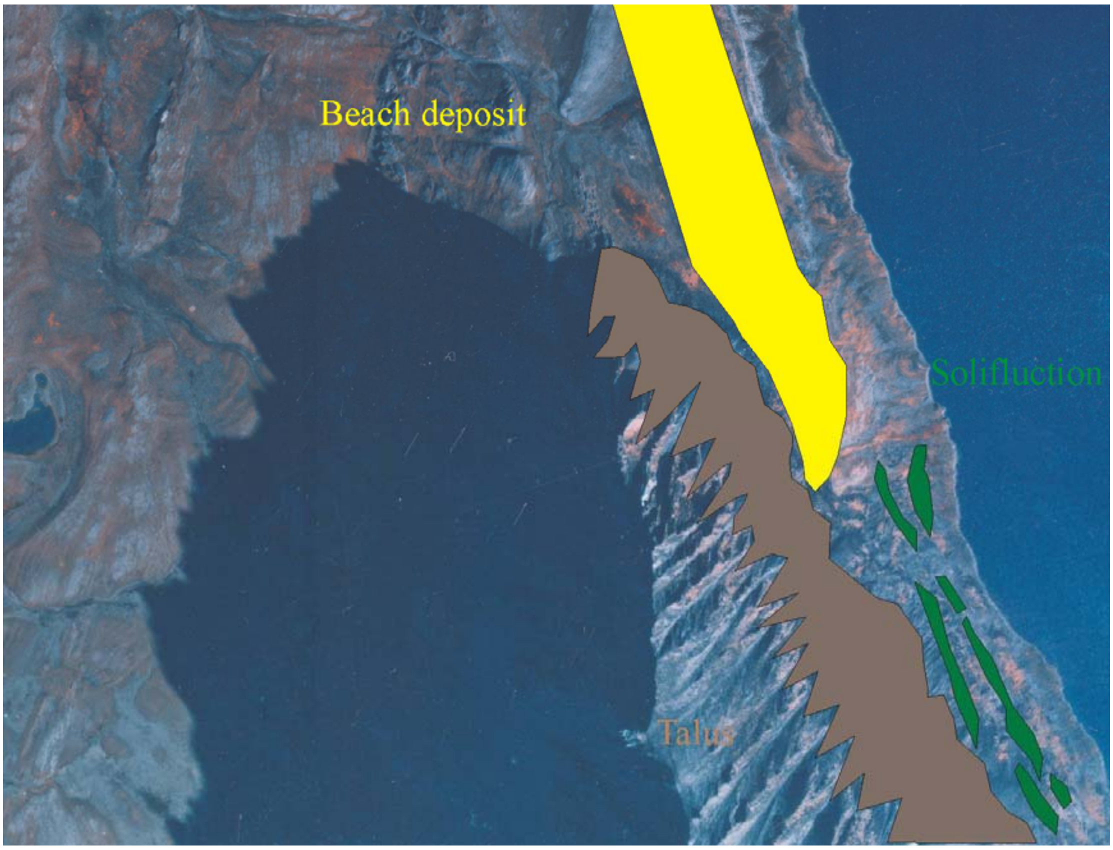


Fig. 3.2.3. Geomorphological map of the eastern side of Griegaksla.
The beach deposit, talus slopes and solifluction areas are indicated.
(Photo: Norsk polarinstitutt.)

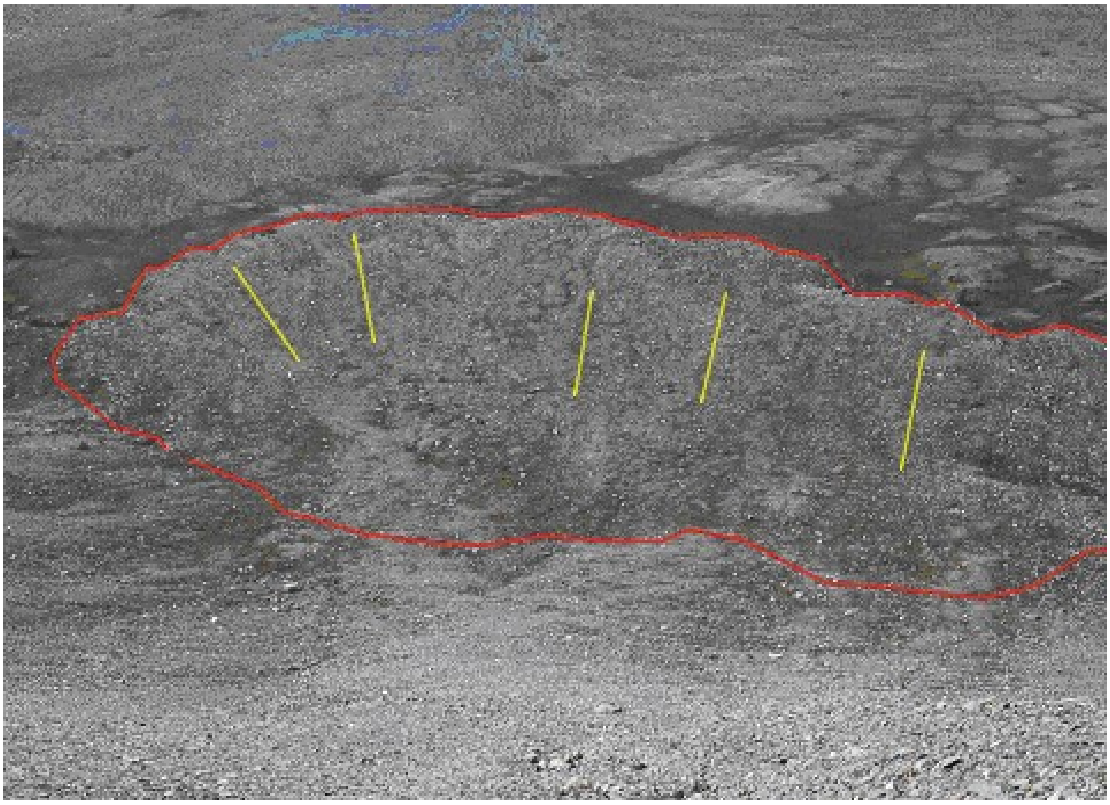


Fig. 3.2.4. The solifluction lobe on the western side of Griegaksla. Flowlines are indicated.
View towards west.

Description of the solifluction

The solifluction phenomena along Griegaksla appear very different on the two sides of the mountain (see location map in figure 3.2.1). On the western side, pronounced solifluction sheets were visible (fig. 3.2.5). The appearance along the eastern side is more chaotic, with unclear lobes floating down the hill as individual streams of soil (fig. 3.2.8). Sometimes the lobe was covered with vegetation, and had a clearer appearance (fig. 3.2.9).

Along Griegaksla, a majority of the solifluction is considered to be turf-banked lobes. However, some of the fronts have mostly big boulders, and should therefore be classified as stone-banked.

West side

On the western side, the solifluction sheets are located below a system of well-developed coalescing talus cones. They can be followed along the mountain sides more than 50 km to the south, only with minor gaps. According to previous investigations done by Åkerman (1996), the lobate fronts of the sheet have an average height of 0.65 m. Due to over-saturation, some of the fronts have collapsed and are not more than 0.2-0.3 m high. The sheet and the lobes are flowing out over flat surfaces of raised beach ridges and marine terraces at an average height of 38 m a.s.l. in the northern part and 75 m a.s.l. in the southern part. Generally, the lobes are freely flowing out, but in some places they are blocked by rock outcrops or dissected by small streams.

The solifluction on the western side consists of sheets with different surface sorting. In the northernmost part of Griegaksla, the solifluction consists of mainly large rocks and boulders, intermixed with finer sediments. Further south the solifluction surface is dominated by smaller rocks and vegetation. The solifluction sheet that is investigated in greater detail in the present paper is ca. 110 m long, 180 m wide and has a frontal riser of ca. 30-80 cm. Figure 3.2.5 gives an overview of the sheet. The surface slope of the sheet is between 7 and 17º. The altitude reached by the terminus of the solifluction sheet is 20 m a.s.l., while the head is at approximately 50 m a.s.l. Some areas of the sheets are very wet, and the water is seen seeping from the downstream side of large boulders lying on the ground (fig. 3.2.6). Figure 3.2.7 shows how the surface of the sheet is sorted in some places.

Fig. 3.2.5. Overview of the sheet on the western side. The trench is marked with an arrow.

Fig. 3.2.6. Picture showing how the water is coming to the surface beneath the boulders.

Fig. 3.2.7. Detail of the lobe on the western side of Griegaksla, showing sorted rocks.

East side

Along the eastern side of Griegaksla, the solifluction looks like small lobes flowing down the mountain side. The borders of the solifluction are unclear, and the dip of the ground varies greatly on a local scale. Still, there are distinct areas where the solifluction appears: Near the beach of lake Linnévatnet (situated at 10 m a.s.l.), the slope angle is about 4º. Here the ground surface is sorted, with stripes of boulders between areas covered by moss. These stripes are perpendicular to the terrain slope. Further up the slope, at 15 m a.s.l. the ground steepens, and the slope angle becomes about 18º. The ground is mostly covered with bigger rocks. At 20 m a.s.l., the slope angle then declines to about 8º, and here the solifluction lobes are found. They creep around and flow over the outcrops. Sorting is also found here, with stripes of bigger boulders between the vegetation. At 40 m a.s.l., there is another edge which is 25º steep. Solifluction lobes are found at this level, between 40-50 m a.s.l. The ground has a gentler slope above the edge, still with solifluction lobes. At about 50 m a.s.l., the lower part of the talus slope begins and the solifluction is no longer visible. Figure 3.2.8 shows an overview of the area as seen from the talus slope above. Figure 3.2.10 shows a drawing made to give a general picture of the solifluction in this area, as described.

Fig. 3.2.8. Overview of the solifluction at the eastern side of Griegaksla, view towards east.
The surface of the solifluction lobes are often covered with vegetation, the front contains large boulders.

Fig. 3.2.9 Solifluction lobe at the eastern side of Griegaksla. View towards SW.

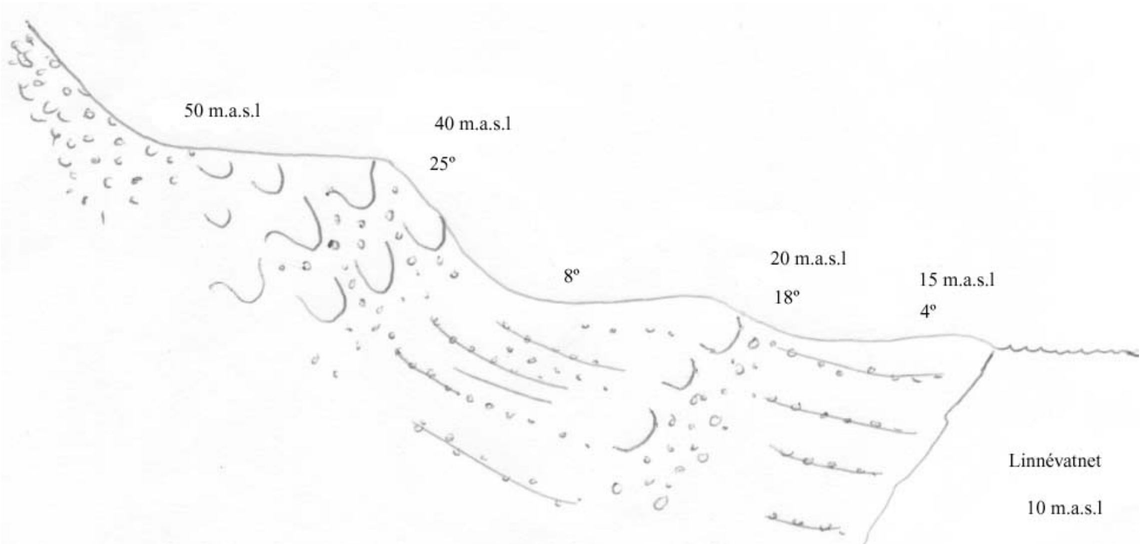


Fig. 3.2.10. Drawing made by the author, showing the general distribution of solifluction on the eastern side of Griegaksla. View towards north.

3.2.4 Methods

Digging a trench

A 7 metre long and 90 cm deep trench was dug into the end of a solifluction sheet on the western side of Griegaksla (fig. 3.2.1 and 3.2.5). The location of the trench was chosen due to the ease of digging and a relatively low frequency of big boulders. A supplementary reason was that the chosen site appeared to be relatively dry; with no water running from the terminus of the lobe. The sheet was 30 cm high at the end, and the surface slope was about 10 degrees. The direction of the trench was 60 º NE, parallel to the local surface slope. In order to reach the bottom of the active layer, the upper end of the trench was dug deeper than 90 cm. At 1.7 m depth the permafrost was reached. After some days, the sediment at the base of the hole thawed, and the hole could be deepened to 2 m. At this depth, a layer of bedrock (phyllite) was reached, making further excavation impossible.

Logging the trench

Most of the 2 metre deep profile consists of a mixture of sand, clay and stones of different sizes. The biggest rocks were about 20 cm in diameter, but most of them were between 2 and 10 cm. 140 cm below the surface, a layer of sorted, rounded rocks of the same size were found. This probably represents the former beach surface formed during the early Holocene regression, now lifted to 30 m a.s.l. The thickness of the strandflat layer was 20 cm, and the mean size of the stones was 2 cm. The layer was well sorted (all the pebbles were about the same size), and consisted of just a small amount of coarse grained sand. Under the strandflat layer, coarse grained sand was found. This layer was 30 cm deep and well sorted. A slide caliper showed that the mean size of these sand grains was 0.3 cm.

A predefined scale of grainsizes was used to distinguish between the different layers in the trench, and a log was made (fig. 3.2.11).

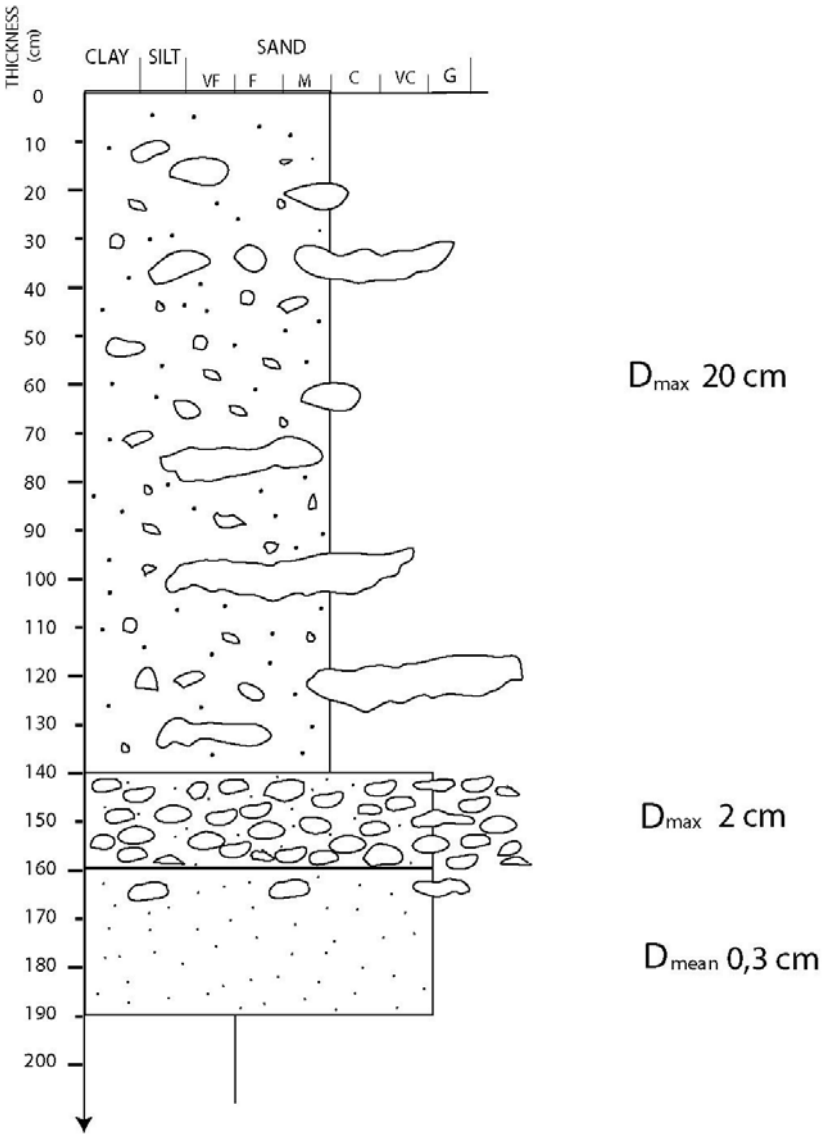


Fig. 3.2.11. A log of the hole dug into the solifluction lobe. (Log made by the author).

A buried vegetation layer was found in the outermost part of the trench, and reflects that the sheet has been overriding the vegetation cover on the ground. The layer could only be traced 50 cm into the trench, further in it was impossible to find the old vegetation. The modern vegetation on the solifluction sheet is discontinuous, which makes a well-defined buried vegetation layer unlikely.

The horizontal profile of the trench was quite uniform. Most of the trench was about 90 cm deep, and only the uppermost layer was visible. There were no visible changes in colour or grainsize between the outermost and the innermost part of the trench.

<u>Fabric analysis</u>

In the trench, 3-4 metres from the edge, fabric analysis was carried out. This was done at three depths: the upper 10 cm, at 40-50 cm depth and at 80-90 cm depth. 25 oblong stones were picked at each depth, measuring the length of the three axes, the orientation and dip of the longest axis of the stone.

<u>Comparing of photos</u>

In 2004, H. Christiansen took pictures of two solifluction fronts nearby the place were the trench was dug. The year after, another picture was taken from the same spot to see if there had been any visible movements.

<u>Future work</u>

Ten fix points were established to use in future work on determining the rate of solifluction movement. The fix points were painted on rocks in front of the solifluction lobe where the trench was dug. By each of these fix points, three rocks on the lobe itself were marked. The distances between the fix point and the marks were measured. Later it will be possible to redo the measurements, and see if the rocks on the lobes have been moving in relation with the fix points.

Before the 2m deep hole at the end of the trench was closed, a vertical standing rope was put into it. The plan is that it will be possible to excavate the rope at a later stage, and estimate the vertical deformation profile by measuring the curvature of the rope. 1.05 m of the rope is visible at the surface, and is situated 9.6 m from fix point number 8.

<u>Schmidt-hammering</u>

The Schmidt hammer is a tool to test the hardness of the rocks. A weathered rock is normally softer than a fresh rock surface; a weathered surface will therefore give a lower value (Aydin and Basu, 2005). Due to this, the hammer can be used to measure the relative age of the stones that are lying on top of the solifluction lobes. This method will give an indication of time, i.e. how long has passed since the stone left the outcrop. In this study, some measurements were done at three different altitudes at the two sides of the Griegaksla. At each altitude, a number of measurements were taken to get an average.

3.2.5 Results

Comparing of photos

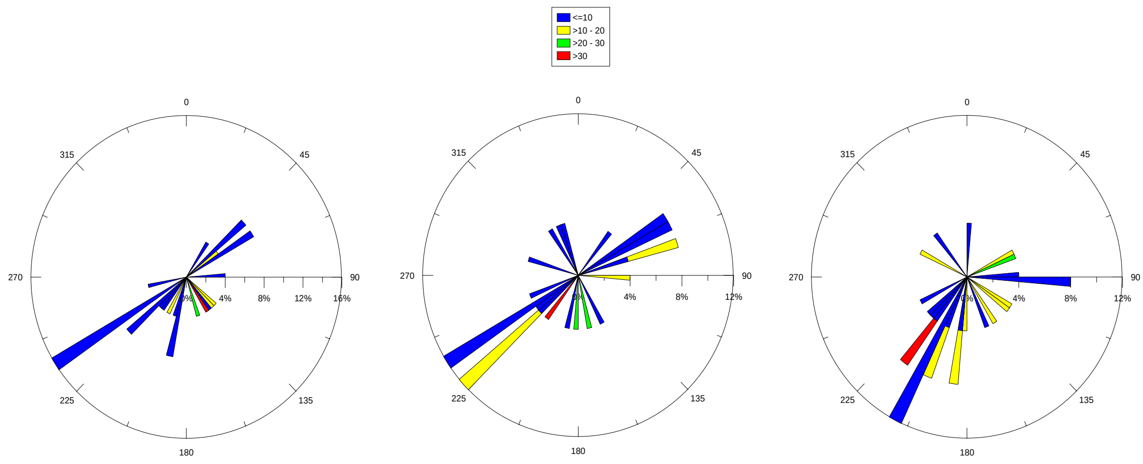
Looking at the two pictures taken of the solifluction lobes during two years, only a little movement can be seen in one of the lobes. This movement is visible in the area of the lobe that has little vegetation and looks sandier. One stone is marked with an arrow; this stone has moved further down on the photo from 2005. This movement might be caused by other factors than solifluction movement, for example by an animal, but is worth noticing. Figure 3.2.12 shows the picture taken in 2004; figure 3.2.13 shows the picture taken in 2005.

Fig. 3.2.12 Picture taken of a solifluction lobe in 2004. Note the stone marked with an arrow.
(Photo: H. Christiansen)

Fig. 3.2.13. Picture of the same spot taken in 2005. The stone marked with an arrow has moved.
(Photo: K. Senger)

Rose diagram

Based on the fabric analysis, three rose diagrams were made (fig. 3.2.14). They show the orientation of the length axes of the oblong stones at three levels in the trench. The colours indicate the dip of the stones.



upper 10 cm middle (40-50 cm depth) lower (80-90 cm depth**)**

Fig. 3.2.14. Rose diagram showing the orientation and dip of the longest axis of the oblong stones in the trench. Blue colour indicates that the dip is equal to or less than 10º. Stones dipping 10-20º are marked with yellow. Green indicates a dip of 20-30º, and red a dip of more than 30º.

The first diagram shows the stones found in the upper 10 cm of the trench. It is clear that the stones have a preferred orientation, either towards NE or SW. 16% of the stones are pointing towards 235-240 º, 32% of the stones are pointing towards 200-240 º. Some stones are also pointing opposite the flow direction, about 30% of them are pointing between 40º and 95º. A majority of the stones dip no more than 10º.

The diagram in the middle shows the stones found between 40 and 50 cm depth. The direction of the stones varies more here, but there is still a preferred direction. About

30% of the stones are pointing towards 220-240º. 25% of the stones are pointing in the opposite direction, towards 45-90º. Only 25% of the stones have a dip of more than 10º.

The last diagram is showing the orientation of the stones found at 80-90 cm depth. The preferred direction of the stones is slightly different than in the uppermost part of the soil. Here, the stones are pointing more towards the SSW. 32% of the stones are pointing towards 210-230º. At this depth, there are only few stones pointing in the opposite direction.

It seems like most of the longest axes of the stones were dipping towards 210-240º. This corresponds well to the direction of the solifluction flow, which was estimated to about 240º.

<u>The Schmidt hammer</u>

In three levels on the solifluction lobes on either side of Griegaksla, several measurements with the Schmidt hammer were undertaken. The mean values are shown in table 3.2.1.

Table 3.2.1 Mean values from the Schmidt hammer used on both sides of Griegaksla.

	Lower part of the solifluction	Middle part of the solifluction	Upper part of the solifluction
Eastern side of Griegaksla	52,33	55,33	53,45
Western side of Griegaksla	54,07	52,00	45,47

At the eastern side, the results of the hammering were almost the same at all three altitudes. This might be because the slope is so chaotic and the stones near the shore are not weathered as much as the stones in the upper part of the hill.

At the western side, the values of the hammering were higher at the bottom of the solifluction lobe than at the top. This does not follow the theory that the stones at the bottom are older and more weathered, and therefore should have a lower value. Why this is so, is unclear. One explanation might be that the boulder used for measuring at the top was an exception, and that an examination on more boulders would give other results.

Lichen studies

Table 3.2.2 Mean size of the lichen on the western side of Griegaksla.

	Lower part of the solifluction	Middle part of the solifluction	Upper part of the solifluction
Mean size of the longest axis (mm)	27.9	29.6	28.3

According to theory (e.g. Beschel-Roland, 1961) the diameter of the lichens will be larger towards the end of a solifluction lobe, since they have been exposed the longest. In the present study, it was the middle part that had the largest lichens, and the lower part the smallest. An explanation to this might be that the stones in the lower part have been exposed to more movement, which prevent lichen growth. Another explanation is that the investigated stones in this study were too few to give good data, and that the measuring method is uncertain.

3.2.6 Discussion

Subsequent to the trench being dug a period of intense rainfall followed. During one night about 12.3 mm of rain was recorded according to the meteorological station at Kapp Linné. This rainfall made the trench collapse, as seen in figure 3.2.15. The major collapse indicates that a rainfall makes the solifluction sheet very unstable. So maybe it is not only the freezing and thawing that makes the earth move downhill, but also periods of rain.

Fig. 3.2.15. The trench before and after the collapse. The pictures are taken with an interval of one night.

The height of the lower part of the solifluction sheet on the western side of Griegaksla is 20 m a.s.l. The uppermost part of the solifluction sheet that was investigated at the western side is 50 m a.s.l. Landvik et al. (1987) have done a study of beach deposits in the same area, and with the help of whalebones and seaweed dated the beach levels. According to them, the shoreline at 50 m a.s.l. is of Younger Dryas age (10,000 years B.P.). This means that the lobe must have started to develop after this time; before the Younger Dryas it was lying under water and could not develop. According to their shoreline displacement curve for the Ytterdalen area (about 30 km south of Kapp Linné), an altitude of 20 metres was a shoreline 8000 C14 years B.P, which means that the end of the lobe must be of this age or younger.

This gives the lobe a minimum of 2000 years to develop, if it stopped moving 8000 years B.P. If it has been moving all the time since 10,000 years B.P., it has used 10,000 years. Using an approximate lobe length of 110 m, this gives an average speed of between 1.1 and 5.5 cm/yr over these time ranges. Papers on Spitsbergen have reported speed rates of 4-6.5 cm/year (Repelewska-Pękalowa and Pękala, 1993), 1.5-3 cm/yr (Büdel, 1960, 1963), 5-12 cm/yr (Jahn, 1967), 3.4-8.1 cm/yr (Åkerman, 2005), 2.5 cm/yr (Matsuoka and Hirakawa, 2000) and 0-40 cm/yr (Repelewska-Pękalowa and Pękala, 1993). The calculated velocity for the solifluction lobe in the present paper is within the common rate for the area.

According to Repelewska-Pękalowa and Pękala (1993), who did studies in the region of southern Bellsund (south of Kapp Linné), the solifluction deposits on Spitsbergen are young. They represent one main cycle of periglacial sedimentation lasting from the late Pleistocene up to the present time. Evidence of this are the solifluction forms developed in slope deposits overlying the marine terraces and at the base of the terrace cliffs. The marine terraces were raised in the terminal phases of the Pleistocene and at the beginning of the Holocene.

The reason why the solifluction looks different on the two sides of the Griegaksla is probably due to the rock type. This is different on the two sides of the mountain. On the eastern side, the boulders are bigger than on the western side. According to the geological map in figure 3.2.2, the eastern side consists of sandstone from Lower Carboniferous, and on the western side, phyllite dominates.

4 Discussions of solifluction

4.1 Discussion of climate and solifluction

Veit et al. (1995) did some studies in Austria, trying to find a relationship between solifluction and climate. They found that the solifluction movement is mainly dependent on the following factors:

- Number of days with snow cover > 1 cm (Oct-Nov) or the first snowfall.
- Duration of the snow cover in winter.
- Number of days with snow > 1 cm in summer, especially in August.
- Summer temperature (June-July).

The two summer parameters are related to each other. When summers are cool, a larger amount of precipitation will fall as snow. Frost depth in the ground is dependent upon when the first snow falls. A late snowfall will give the ground time to freeze. The temperature during winter is of secondary importance, because the snow isolates the ground underneath. Snow in summertime and low summer temperatures protects the frost in the ground, and leads to a slow thaw which gives greater movements (Veit et al. 1995).

Since solifluction is dependent so many factors, it is not easy to reconstruct climate based on solifluction movement. High activity can be caused by coolness, winters with little snow and/or cool summers (Veit et al. 1995).

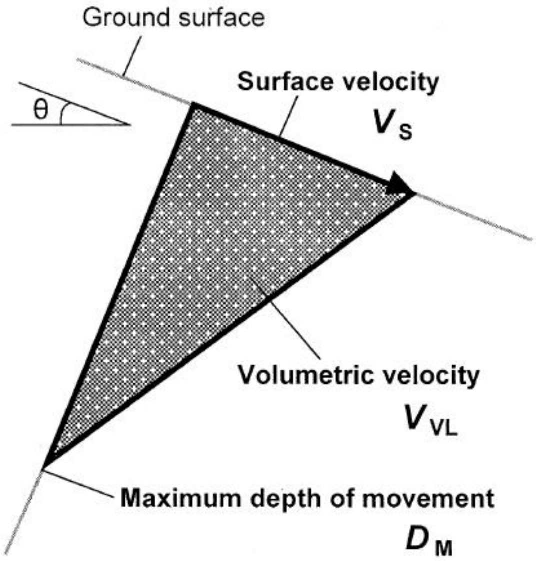


Fig. 4.1.1. Soil movement parameters (Matsuoka, 2001).

According to Matsuoka (2001), three parameters define downslope movement: the surface velocity, volumetric velocity and maximum depth of movement (fig. 4.1.1). These parameters vary greatly with MAAT, and this is illustrated by Matsuoka in figure 4.1.2. A data point in this figure represents the mean value for a study site, except where data are summarized for each landform type.

In places with mostly cold and continuous permafrost and a MAAT < -6ºC, the surface velocity is very low, rarely more than 5 cm/yr. When the MAAT increases, the surface velocity rises. The maximum speed is reached with a MAAT between -3ºC and -5ºC; within this zone the speed often exceeds 10 cm/yr. Both slopes underlain by warm permafrost (temperature close to 0ºC) and those lacking permafrost are represented in this zone. However, high surface speed values are unlikely to result from the presence of permafrost or deep seasonal frost. A surface velocity of more than 10 cm/yr is usually associated with a high frequency of diurnal freeze-thaw cycles which normally occur on mid- to low-latitude high mountains (Matsuoka, 2001).

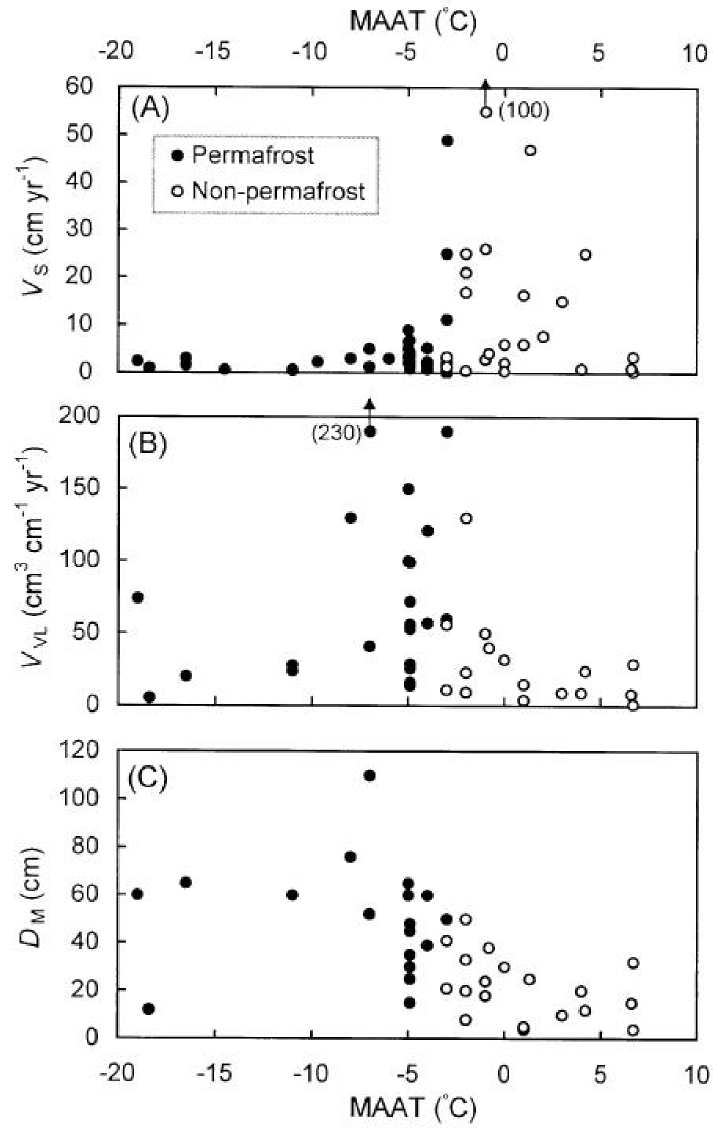


Fig. 4.1.2. Periglacial soil movement as a function of the mean annual air temperature (MAAT) See fig. 4.1.1 for definitions of V_S , V_{VL} and D_M (Matsuoka, 2001).

Matsuoka and Hirakawa (2000) found that the deepest movement was 110 cm, recorded in Svalbard. In non-permafrost areas, the depth ranges from a few centimetres to 50 cm (Matsuoka, 2001). The volumetric velocity reaches a maximum in the warm permafrost zone, often in excess of 100 cm³/cm/yr. The velocity decreases towards both the cold permafrost and seasonal frost zones. This indicates that solifluction produces the most rapid denudation where slopes are underlain by warm permafrost.

4.2 Discussion of topography and solifluction

Several mathematical models have been suggested to describe the dependence of movement on inclination. The simplest model, as outlined below by Matsuoka (2001), gives the surface velocity Vs by

$$V_S = H_F \tan\theta \qquad (1)$$

where H_F is the heave amount perpendicular to the slope surface and θ is the slope angle (e.g., Williams and Smith, 1989). For slopes subject to multiple frost heave cycles, Eq. (1) is rewritten as:

$$V_S = \sum H_F \tan\theta \qquad (2)$$

where V_S gives the annual surface velocity and $\sum H_F$ the annual total heave amount. V_S in Eqs. (1) and (2) is also called the potential frost creep. In reality, soil cohesion yields some retrograde movement that reduces V_S (Washburn, 1979). Where needle ice activity prevails, rotational movement or toppling of ice needles intensifies the effect of inclination such that:

$$V_S = C \sum H_F \tan^2\theta \qquad (3)$$

where C is a constant (Higashi and Corte, 1971; Matsuoka, 1998b)

From some polar mountains, such as in Svalbard (Åkerman, 1996; Sawaguchi, 1995; cf Matsuoka, 2001) and the Canadian Arctic (Washburn, 1967) there has been reported increasing solifluction rates with inclination. In the lower-latitude alpine areas, the influence of inclination is often masked by the spatial variation in other factors like freeze-thaw frequency, surface texture and moisture distribution (e.g. Benedict, 1970; Harris, 1981) although Pérez (1987) found a rough correlation between rate and inclination on slopes dominated by needle ice creep.

Matsuoka (2001) has plotted available field data in an inclination-velocity diagram (fig. 4.2.1). Each point in this diagram represents data for an individual plot. Because Eqs. (2) and (3) suggest different responses to inclination, two groups of sites should be segregated: those governed by diurnal freeze-thaw action (where needle ice creep may prevail) and the rest. Only a weak correlation is found between Vs and tan θ in both groups (fig. 4.2.1A). However, apart from the sites dominated by diurnal freeze-thaw action, there appears to be a maximum Vs for a given inclination, approximated by the line:

$$V_S = 100 \tan\theta$$

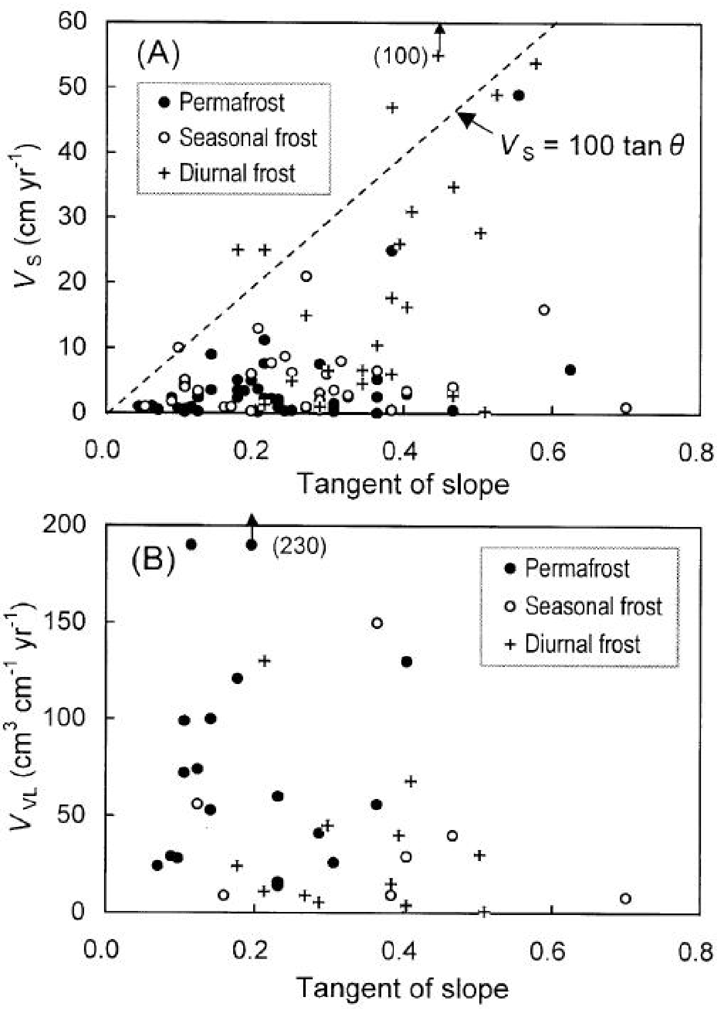


Fig. 4.2.1. Solifluction rates as a function of the slope gradient.
See fig. 4.1.1 for definitions of V_S and V_{VL} (Matsuoka, 2001).

The connection between inclination and volumetric velocity is not clear (fig. 4.2.1B), but there seems to be a weak trend in which velocity decreases with inclination. This trend is possibly controlled by the thickness of frost-susceptible fine debris. Steep slopes cannot hold a thick layer of fine debris that is easily removed, whereas downslope declination allows sedimentation of the fine debris on the lower slopes. As a result of this, maximum depth of movement increases downslope, possibly aided by a rise in moisture availability. The volumetric velocity will increase despite declination. As stated by Matsuoka (2001), this suggests that the rate and process of solifluction is related indirectly to slope form and inclination as these in turn affect debris thickness, snow distribution, drainage and microclimate.

4.3 Discussion of climate changes and solifluction

Climate changes influences solifluction by controlling the thermal and moisture regimes near the ground surface (e.g., Woo et al., 1992). Long-term cooling or warming can change the ground freezing condition, which will affect solifluction in several ways. One case is when the climate is cooling, and newly produced permafrost intensifies seasonal frost heaving (Matsuoka, 2001). The formation of the impermeable layer will improve moisture availability in the active layer, and this possibly increases the rate of annual frost creep or gelifluction on thawing. However, Matsuoka (2001) points out that permafrost formation does not necessarily enhance moisture availability or solifluction rates. This is for instance the case where slope convexity allows free drainage over permafrost, or where permafrost occurs in the bedrock beneath the regolith.

If the climate is cooling in a place with warm permafrost, two-sided freezing may replace one-sided freezing. This happens where the whole active layer consists of fine-textured soil and poor drainage supports high moisture content of the active layer during summer. Two-sided freezing redistributes ice lenses near the bottom of the active layer. On thawing, plug-like flow may increase both the maximum depth of movement and the volumetric velocity (Woo et al., 1992). However, no changes in solifluction processes may take place if the cooling of permafrost is not accompanied by the occurrence of two-sided freezing.

A warm summer in the permafrost zone may induce partial thawing if ice-rich permafrost, followed by a temporary deepening and acceleration of soil movement. Some large solifluction lobes could have developed by such episodic events (Benedict, 1970).

When the climate gets warmer, the vegetation cover will be influenced. A ground surface that used to be bare may become vegetated, and the binding and thermal insulating effects will influence the solifluction. More vegetation will decrease the maximum annual frost depth in a seasonal frost environment. At the same time, a decrease in the thaw depth may also occur in a permafrost environment, despite partially being offset by rising air temperatures in summer. Such a decline in the freeze-thaw depth may also reduce the volumetric velocity and the maximum depth of movement (Matsuoka, 2001).

Many high mountain slopes with a thin vegetation layer experience high freeze-thaw frequency. A warmer climate will thicken this vegetation layer, and inhibit the diurnal freeze-thaw action which occurs most frequently within the uppermost decimetre of soil (Matsuoka, 2001).

If the amount of precipitation changes, this will affect the soil moisture regime and later the solifluction. A greater rainfall in autumn enhances solifluction (Matsuoka, 2001). Snow has different effects on solifluction. On one hand it insulates the ground like vegetation (French, 1996), on the other hand, late laying snow patches provide meltwater to the seasonally thawing ground. This possibly increases the mobility of the soil, and promotes gelifluction (Harris, 1972; Smith, 1988). At some sites the meltwater maintains a high groundwater level until autumn. This intensifies the diurnal freeze-thaw action and increases the amount of seasonal frost heave (Benedict, 1970). Due to this, frost creep (both diurnal and annual) and gelifluction are probably accelerated (Matsuoka, 2001).

4.4 Discussion of solifluction as a palaeoclimate indicator

Solifluction can indicate climatic fluctuations and trends, and therefore represent a potentially useful source of paleoclimatic information (Åkerman, 2005). Solifluction movements are only possible during periods in which the upper part of the soil is water-saturated. Due to this, the upper part is easily translocated, while the deeper layers are still frozen (permafrost). Fossil solifluction covers are therefore a useful palaeogeographic indicator for reconstructing the chronology of solifluction events by providing evidence of seasonal or constant warming. When recovering the Holocene climatic record, studies on solifluction process dynamics under the periglacial conditions of today can be very helpful (Repelewska-Pękalowa and Pękala, 1993).

By dating the vegetation layer it is possible to indicate the movement rate through time. Several datings have to be done from the beginning of the solifluction lobe until the end. When the distances between the datings are measured, an estimate of the speed is possible.

4.5 Discussion of solifluction on Svalbard and Dovre

The two field sites studied in the present paper are different when it comes to climate, vegetation and ground conditions. A comparison between the solifluction features and the mentioned factors is therefore interesting.

The vegetation cover is different at the two field sites. At Dovre, the vegetation is thick and covers the whole area. In Svalbard, the vegetation is sparse and thin. Most of the scientists assume that vegetation retards solifluction movement. If this is the case, the velocity should be higher in Svalbard than at Dovre.

Permafrost favours solifluction, and on Svalbard there is most certainly continuous permafrost. At Dovre, there is possible sporadic permafrost, this is evident from the palsas visible within the vicinity which are known to exist only in permafrost areas.

Low summer temperatures protect the frost in the ground. This therefore leads to a slow thaw and thus great movements (Veit et al., 1995). In Svalbard, the temperatures are generally lower than at Dovre, therefore suggesting greater movements. The MAAT at Kapp Linné, Svalbard is -4.8ºC. According to Matsuoka (2001) maximum surface velocity will occur with a MAAT between -3ºC and -5ºC, which is the case on Svalbard. At Dovre the MAAT is -0.7ºC. According to Matsuokas diagram in fig. 4.1.2, a MAAT of around 0ºC produces great variations in surface speed, normally higher than in the cold permafrost areas. Volumetric velocity and depth of movement is usually lower at -0.7 than at -4.8, according to fig. 4.1.2. This means that these values are probably higher in Svalbard than at Dovre. In addition, previous scientists have recorded the deepest movement (D_M = 110 cm) in Svalbard (Matsuoka and Moriwaki, 1992), so this is probably the case.

Fabric analysis carried out on Svalbard and Dovre shows that the oblong stones prefer to point the longest axis towards the flow direction. The trend is clearer at Dovre than in Svalbard, but is in both cases quite convincing. In the upper and middle level at both Svalbard and Dovre, a number of the stones point opposite the flow direction. The reason why this trend is absent in the lower levels might be that the weight and movement of the layer above has forced the stones to dip downwards, in the same direction as the flow. At Dovre it seems like the flow direction has been stable towards 315º since the beginning of the lobe deposition. At Svalbard, the direction of the oblong stones varies slightly when comparing the three levels of analysed stones. At the lowest level, the stones are pointing towards approximately 210º. At the middle and upper level, the stones are mostly dipping towards about 235º. This change is probably due to a change in flow direction.

5 Conclusions

Measurements of the orientation of the oblong stones in solifluction lobes on Svalbard and Dovre show that most of the stones points towards the same direction. This direction corresponds well to the present flow direction of the solifluction lobes.

The soil in the solifluction sheets is quite homogenous in both field sites.

Due to the palsas nearby, and the widespread solifluction in the field area, it is likely that there is or has been permafrost at Dovre.

The collapse of the trench in the solifluction lobe shows that the soil gets very unstable during rainfall. This indicates that rainfall might have a big influence on the movement of the solifluction lobes.

Solifluction occurs in places where the ground is seasonal or perennially frozen. Being a slow process, the big extension of solifluction on both sides of Griegaksla tells us that the climate must have been cold for a long time, at least since the Younger Dryas. In Younger Dryas, the sea level was 50 meters above the present sea level. The solifluction at Griegaksla, which is situated at 10-50 m a.s.l., must therefore have started to develop after this, since it doesn’t develop under water.

6 References

Andersson, J.G., 1906. Solifluction, a component of subaerial denudation. *Journal of Geology* 14, 91-112.

Aydin, A. and Basu, A., 2005. The Schmidt hammer in rock material characterization. *Engineering Geology,* 81 (1), 1-14

Ballantyne, C.K. and Harris, C., 1994. *The Periglaciation of Great Britain.* Cambridge University Press, Cambridge. 330 pp.

Baulig, H., 1956. Pénéplaines et pédiplaines. *Soc. Belge etudes géog.* 25 (1), 25-58.

Benedict, J.B., 1970. Downslope soil movement in a Colorado alpine region: rates, processes, and climatic significance. *Arctic Alpine Research,* 2 (3), 165-183, 219-226.

Benedict, J.B., 1976. Frost Creep and Gelifluction Features: A Review. *Quaternary Research* 6, 55-76.

Berthling, I., 2001. Slow periglacial mass wasting-processes and geomorphological impact. Case studies from Finse, southern Norway and Prins Karls Forland, Svalbard. *PhD thesis.* University of Oslo.

Beschel-Roland, E., 1961. Dating rock surfaces by lichen growth and its application to glaciology and physiography (lichenometry). In: Raasch, G.O. (Ed.), *Geology of the Arctic*. Vol. 2. Toronto: University of Toronto Press, 1044-1062.

Büdel, J., 1960. Die Frostschutt-Zone Spitzbergens. *Coll. Geogr.* 6, 105 pp.

Büdel, J., 1963. Klimagenetische Geomorphologie. *Geogr. Rundschau* 15, 269-285.

Denton, G.H. and Karlén, W., 1973. Holocene climatic variations-their pattern and possible cause. *Quaternary research* 3, 155-205.

Dylik, J., 1951. Some periglacial structures in Pleistocene deposits of Middle Poland. *Bulletin de la Societe des Sciences et des letters de lodz.* 3, 1-6.

Egginton, P.A. and French, H.M., 1985. Solifluction and related processes, eastern Banks Island, N.W.T. *Canadian Journal of Earth Sciences,* 22, 1671-1678.

Frakes, L.A., 1979. *Climates throughout geologic time*. Elsevier, Amsterdam. 310 pp.

Francou, B. and Bertran, P., 1997. A multivariate analysis of clast displacement rates on stone-banked sheets, Cordillera Real, Bolivia. *Permafrost and Periglacial Processes,* 8, 371-382.

French, H.M., 1974. Mass-wasting at Sachs Harbour, Banks Island, N.W.T., Canada. *Arctic and Alpine Research,* 6, 71-78.

French, H.M., 1976. *The Periglacial Environment.* Longman, London. 309 pp.

French, H.M., 1996. *The Periglacial Environment*, 2nd ed. Longman, Essex. 341 pp.

Gamper, M., 1983. Controls and rates of movement of solifluction lobes in the Eastern Swiss Alps. In: *Proceedings, 4th International Conference on Permafrost*, Vol. 1. Washington, DC: National Academy Press, 328-333.

Harris, C., 1972. Processes of soil movement in turf-banked solifluction lobes, Okstindan, northern Norway. In: *Polar Geomorphology.* Institute of British Geographers Special Publication, No. 4, 155-174.

Harris, C., 1973. Some factors affecting the rates and processes of periglacial mass movements: *Geografiska Annaler* 55A, 24-28.

Harris, C., 1981. Periglacial Mass-Wasting: A Review of Research. BGRG Research Monograph, Vol. 4, *Geo Abstracts*, Norwich. 204 pp.

Harris, C., 1989. Mechanisms of mass movement in periglacial environments. In: Anderson, M.G., Richards, K.S. (Eds.), *Slope Stability*. Wiley, Chichester, 531-559.

Harris, C. 1993. The role of climate and soil properties in periglacial solifluction: evidence from laboratory simulation studies. In: Frenzel, B. (Ed.), *Special Issue 6 European Science Foundation Project on European Paleaoclimate and Man,* Akademie der Wissenschaften und Literature Mainz, Gustav Fischer Verlag, Stuttgart, 295-398.

Harris, C., 1996. Physical Modelling of Periglacial Solifluction: Review and Future Strategy. *Permafrost and Periglacial Processes*, 7, 349-360.

Harris, C., Davies, M.c.R. and Coutard, J.P. 1996. An experimental design for laboratory simulation of periglacial solifluction processes. 21, 67-75.

Higashi, A. and Corte, A.E., 1971. Solifluction: a model experiment. *Science,* 171, 480-482.

Holdgate, M.W., Allen, S.E. and Chambers, M.J.G. 1967. A preliminary investigation of the soils of Signy Island, South Orkney Islands. *British Antarctic Survey Bulletin,* 12, 53-71.

Holtedahl, O. (Ed.), 1960. *Geology of Norway.* Norges geologiske undersøkelse nr. 208. Aschehoug, Oslo. 540 pp.

Humlum, O., Instanes, A. and Sollid, J.L. 2003. Permafrost in Svalbard: A review of research history, climatic background and engineering challenges. *Polar Research* 22 (2), 191-215.

Huntley, B., and Prentice.C., 1988. July temperatures in Europe from pollen data 6000 years before present. *Science,* 241, 687-690.

Jahn, A., 1960. Some remarks on evolution of slopes on Spitsbergen. *Zeitschrift für Geomorphologie, Supplementband 1,* 49-58.

Jahn, A., 1961. Quantitative analysis of some periglacial processes in Spitsbergen. *Nauka O Ziemi II, Seria B,* 5, 3-34.

Jahn, A., 1967. Some features of mass movement on Spitsbergen slopes. *Geografiska Annaler,* 49A, 213-225.

Kirkby, M.J., 1995. A model for variations in gelifluction rates with temperature and topography: implications for global change. *Geografiska Annaler,* 77A, 269-278.

Landvik, J.Y., Mangerud, J. and Salvigsen, O., 1987. The Late Weichselian and Holocene shoreline displacement on the west-central coast of Svalbard. *Polar Research 5 n.s.*, 29-44.

Lewkowicz, A.G., 1988. Slope processes. In: Clark, M.J. (Ed.) *Advances in Periglacial Geomorphology.* Wiley, Chichester, 325-368.

Liestøl, O. 1977. *Pingos, Springs and Permafrost in Spitsbergen.* Norsk Polarinstitutt Årbok 1975.

Mackay, J.R., 1981. Active layer slope movement in a continuous permafrost environment, Garry Island, Norhwest Territories, Canada. *Canadian Journal of Earth Sciences,* 18, 1666-1680.

Matsuoka, N., 1994. Continous recording of frost heave and creep on a Japanese alpine slope. *Arctic and Alpine Research,* 26, 245-254.

Matsuoka, N., 1996. Soil moisture variability in relation to diurnal frost heaving on Japanese high mountain slopes. *Permafrost and Periglacial Processes,* 7, 139-151.

Matsuoka, N., 1998a. The relationship between frost heave and downslope soil movement: field measurements in the Japanese Alps. *Permafrost and Periglacial Processes,* 9, 121-133.

Matsuoka, N., 1998b. Modelling frost creep rates in an alpine environment. *Permafrost and Periglacial Processes,* 9, 397-409.

Matsuoka, N., 2001. Solifluction rates, processes and landforms: a global review. *Earth-Science Reviews,* 55, 107-134.

Matsuoka, N. and Moriwaki, K., 1992. Frost heave and creep in the Sør Rondane Mountains, Antrarctica. *Arctic and Alpine Research,* 24, 271-280.

Matsuoka, N., Hirakawa, K., Watanabe, T. and Moriwaki, K., 1997. Monitoring of periglacial slope processes in the Swiss Alps: the first two years of frost shattering, heave and creep. *Permafrost and Periglacial Processes,* 8, 157-179.

Pérez, F.L., 1987. Downslope stone transport by needle ice in a High Andean area (Venezuela). *Rev. Geomorphol. Dyn.* 36, 33-51.

Pissart, A., 1964. Vitesses des mouvements du sol au Chambeyron (Basse Alpes). *Biuletyn Peryglacjalny,* 14, 303-309.

Price, L.W., 1973. Rates of mass wasting in the Ruby Range, Yukon Territory. In: Permafrost; *North American Contribution, Second International Permafrost Conference, Yakutsk, USSR.* Washington, DC: National Academy of Science, Publication 2115, pp. 235-245.

Rapp, A., 1960. Recent developments of mountain slopes in Kärkvagge and surroundings, northern Sweden. *Geografiska Annaler,* 42, 71-200.

Rapp, A., 1962. Kärkevagge: some recordings of mass movements in the northern Scandinavian mountains. *Biuletyn Peeryglacjalny,* 11, 287-309.

Repelewska-Pękalowa, J. and Pękala, K., 1993. The influence of local factors on solifluction rates, Spitsbergen, Svalbard. In: Frenzel, B., Matthews, J. A. and Glaser, B. (Eds.), *Solifluction and climatic variation in the Holocene.* Gustav Fischer Verlag, Stuttgart. 251-266.

Rudberg, S., 1962. A report on some field observations concerning periglacial geomorphology and mass movents on slopes in Sweden. *Biuletyn Peryglacjalny,* 11, 311-323.

Rudberg, S., 1964. Slow mass movement processes and slope development in the Norra Storfjäll area, southern Swedish Lappland. *Zeitschrift für Geomorphologie, Supplementband* 5, 192-203.

Sawaguchi, S., 1995. Rates and processes of mass movement on periglacial rubble slopes in Spitsbergen (in Japanese). *Journal of Geography,* 104, 874-894.

Sigafoos, R.S. and Hopkins, D.M., 1952. Soil instability on slopes in regions of perennially-forzen ground (Alaska). In: *Frost actions in soils: A symposium. National Academy of Sciences- National Research Council, Highway Research Board, Special Report No. 2,* 176-192.

Smith, D.J., 1987. Solifluction in the Southern Canadian Rockies. *The Canadian Geographer,* 31 (4), 309-318.

Smith, D.J., 1988. Rates and controls of soil movement on a solifluction slope in the Mount Rae area, Canadian Rocky Mountains. *Zeitschrift für Geomorphologie. Neue Folge, Supplementband* 71, 25-44.

Veit, H., Stingl, H., Emmerich, K.-H. and John, B., 1995. Zeitliche und räumliche Variabilität solifluidaler Prozesse und ihre Ursachen. *Zeitschrift für Geomorphologie. Neue Folge, Supplementband* 99, 107-122.

Washburn, A.L., 1967. Instumental observations of mass-wasting in the Mesters Vig district of North East Greenland. *Meddr. Grønland,* 166, 1-297.

Washburn, A.L., 1979. *Geocryology*: A Survey of Periglacial Processes and Environments. Edward Arnold, London.

Williams, P.J., 1957a. Some investigations into solifluction features in Norway. *Geog. Jour.,* 123, 42-58.

Williams, P.J., 1957b. The direct recording of solifluction movements. *Amer. J. Sci.,* 255, 705-715.

Williams, P.J., 1959. An investigation into processes occurring in solifluction. *Amer. J. Sci.,* 257, 481-490.

Williams, P.J. and Smith, M.W., 1989. *The Frozen Earth. Fundamentals of geocryology.* Cambridge University press, Glasgow. 306 pp.

Woo, M., Lewkowicz, A.G. and Rouse, W.R., 1992. Response of the Canadian permafrost environment to climate change. *Phys. Geogr.* 13, 287-317.

Åkerman, H.J., 1980. Studies on periglacial geomorphology in West Spitsbergen. *PhD Thesis.* Lunds Universitets geografiska institution avhandlingar LXXXIX.

Åkerman, H.J., 1993. Solifluction and creep rates 1972-1991, Kapp Linné, West Spitsbergen. In: Frenzel, B., Matthews, J.A. and Glaser, B. (Eds.), *Solifluction and climatic variation in the Holocene.* Gustav Fischer Verlag, Stuttgart. 225-250.

Åkerman, H.J., 1996. Slow mass movements and climatic relationships, 1972-1994, Kapp Linné, West Spitsbergen. In: Anderson, M.G. and Brooks, S.M. (Eds.), *Advances in Hillslope Proccesses, Vol. 2.* Wiley, Chichester, 1219-1256.

Åkerman, H.J., 2005. Relations between slow slope processes and active-layer thickness 1972-2002, Kapp Linné, Svalbard. *Norsk Geografisk Tidsskrift - Norwegian Journal of Geography* 59, 116-128.

References from the internet:

PhD project (VISTA) at the department of botany, University of Bergen.
URL: http://www.uib.no/bot/qeprg/dovre.htm (date: 24.11.05)

7 Appendices

Appendix 1: Fabric analysis, Dovre
Appendix 2: Fabric analysis, Svalbard
Appendix 3: Schmidt hammer values
Appendix 4: Established fixpoints, Svalbard

Appendix 1: Fabric analysis, Dovre

Upper 10 cm

Stone number	Strike (degrees)	Dip (degrees)	A (cm)	B (cm)	C (cm)
1	314	20	32,0	17	8,0
2	314	5	3,1	2,7	1,4
3	312	12	2,5	1,2	1,0
4	130	8	8,2	3,6	2,7
5	100	5	6,5	8,3	2,4
6	345	8	3,5	1,9	1,0
7	332	15	3,7	1,7	1,0
8	308	12	4,6	1,8	1,0
9	320	1	8,7	4,8	2,6
10	320	4	3,3	1,8	0,9
11	310	5	4,6	1,7	1,3
12	160	18	6,0	3,8	2,6
13	138	1	7,1	4,8	2,0
14	360	3	3,0	1,7	0,7
15	314	8	3,9	1,1	0,9
16	120	4	3,7	2,1	1,8
17	350	8	2,7	2,1	1,1
18	150	5	6,2	2,8	1,6
19	130	2	8,1	3	1,6
20	138	5	4,1	1,9	1,8
21	138	2	5,3	2,6	1,6
22	134	1	4,4	2,2	1,5
23	132	5	2,4	1,2	0,8
24	110	5	5,9	3,8	1,3
25	135	2	2,5	0,8	0,7

40-50 cm depth

Stone number	Strike (degrees)	Dip (degrees)	A (cm)	B (cm)	C (cm)
1	330	4	2,8	1,7	1,1
2	312	5	2,4	1,8	1,0
3	315	2	5,4	2,9	1,7
4	313	4	7,3	4,2	1,5
5	140	1	2,3	1,0	0,7
6	135	1	1,8	1,2	0,8
7	340	10	5,0	3,0	1,7
8	308	5	2,6	1,3	1,0
9	320	1	6,9	4,3	2,8
10	160	1	1,3	0,6	0,5
11	328	5	4,0	2,3	1,4
12	146	1	1,7	1,2	0,8
13	102	5	2,5	1,7	1,0
14	320	8	2,9	1,4	0,6
15	340	1	3,4	2,7	1,5
16	330	5	2,3	1,5	0,8
17	305	5	3,0	1,7	1,1
18	320	12	2,0	1,5	0,5
19	340	15	2,1	1,4	0,7
20	325	5	3,5	1,3	0,9
21	240	2	1,8	0,7	0,4
22	165	7	3,6	2,3	1,0
23	166	2	1,3	0,6	0,3
24	348	4	2,4	1,8	1,2
25	320	2	4,0	2,5	0,8

80-90 cm depth

Stone number	Strike (degrees)	Dip (degrees)	A (cm)	B (cm)	C (cm)
1	325	5	2,9	1,6	0,6
2	320	4	7,4	5,3	3,0
3	318	12	6,9	3,6	2,4
4	325	4	1,7	0,8	0,5
5	312	4	1,9	0,9	0,5
6	312	2	2,5	1,8	1,6
7	293	1	2,1	1,7	1,1
8	162	2	2,1	1,1	0,4
9	316	5	2,8	1,4	0,9
10	320	10	5,0	3,6	2,0
11	312	12	2,8	2,1	1,1
12	156	1	2,2	1,2	0,9
13	323	4	4,2	3,0	1,7
14	160	12	3,0	2,3	1,3
15	314	8	9,3	7,1	4,5
16	317	20	2,5	1,4	0,9
17	315	10	3,8	2,0	0,9
18	312	20	4,2	2,3	1,7
19	127	4	5,5	3,3	1,3
20	160	1	2,4	1,7	0,9
21	55	2	4,3	2,6	1,2
22	325	1	6,5	3,8	1,7
23	318	4	5,7	3,8	0,9
24	280	2	1,9	1,3	0,8
25	285	10	3,4	1,7	1,3

Appendix 2: Fabric analysis, Svalbard

Upper 10 cm

Stone number	Strike (degrees)	Dip (degrees)	A (cm)	B (cm)	C (cm)
1	238	0	5,5	3,5	1,3
2	168	28	8,4	6,4	1,7
3	72	12	5,9	4,9	0,9
4	185	24	5,7	2,3	1,1
5	75	10	5,6	3,1	0,4
6	342	8	7,3	4,8	3,9
7	65	4	5,4	3,5	1,8
8	154	6	4,7	3,6	0,8
9	236	5	4,7	2,7	1,5
10	326	2	5,4	2,5	0,9
11	216	54	5,4	2,4	0,6
12	338	3	3,8	2,8	0,4
13	238	4	4,7	2,7	0,8
14	226	4	4,2	1,8	0,6
15	58	4	2,8	1,7	0,6
16	65	8	5,0	3,3	1,7
17	230	12	2,7	2,5	0,6
18	247	2	4,8	3,4	0,7
19	191	3	3,8	1,2	0,5
20	286	5	4,5	3,5	0,7
21	58	8	6,4	2,7	2,0
22	40	3	5,6	2,5	1,5
23	95	12	3,9	2,0	1,1
24	235	4	4,5	2,6	0,7
25	230	15	5,2	2,0	0,8

40-50 cm depth

Stone number	Strike (degrees)	Dip (degrees)	A (cm)	B (cm)	C (cm)
1	240	2	5,5	4,1	1,0
2	228	5	2,8	1,2	0,4
3	210	15	3,3	1,5	0,7
4	195	3	4,1	2,3	0,5
5	225	5	6,0	3,3	0,7
6	238	4	3,4	2,2	0,4
7	240	1	3,4	2,4	0,4
8	220	5	2,9	1,7	0,6
9	240	4	2,5	1,7	0,4
10	46	8	2,9	1,8	0,8
11	146	35	4,2	2,0	0,9
12	58	2	2,7	1,4	0,6
13	164	24	2,6	1,5	0,4
14	56	2	3,2	2,0	0,6
15	50	8	7,0	4,0	1,8
16	197	4	3,4	1,7	0,8
17	34	4	7,0	4,7	2,0
18	137	13	4,9	2,6	1,2
19	52	18	3,4	2,1	0,6
20	230	8	4,8	3,0	1,0
21	260	4	2,8	1,7	0,3
22	145	2	5,4	3,8	0,8
23	134	16	3,9	2,9	0,8
24	89	8	3,9	2,9	0,9
25	192	4	4,2	2,5	1,2

80-90 cm depth

Stone number	Strike (degrees)	Dip (degrees)	A (cm)	B (cm)	C (cm)
1	95	4	4,1	2,4	1,1
2	210	6	6,3	3,6	1,5
3	124	13	3,6	2,6	0,5
4	190	4	4,3	2,1	1,0
5	202	8	4,5	2,3	0,8
6	190	15	6,2	4,2	2,3
7	94	6	4,0	2,1	0,7
8	216	6	2,7	0,9	0,7
9	325	2	3,8	2,7	1,3
10	2	2	5,1	3,7	1,4
11	210	4	2,5	1,7	0,4
12	90	7	2,8	1,7	0,8
13	130	13	4,3	2,3	0,9
14	205	18	4,4	2,7	1,7
15	296	14	2,5	1,2	0,4
16	184	12	3,9	1,6	1,2
17	241	8	3,0	1,2	0,9
18	156	8	2,4	1,3	0,6
19	68	25	2,2	1,3	0,6
20	148	12	2,7	1,2	0,4
21	220	32	3,0	1,4	1,4
22	225	5	3,0	2,2	0,9
23	230	2	3,0	1,5	0,9
24	65	18	4,8	2,4	1,7
25	210	5	3,8	1,9	0,7

Appendix 3: Schmidt hammer values

West side of Griegaksla:

	Lower end of sheet, 30 m.a.s.l	Middle	Top of sheet, 44 m.a.s.l
	51	58	46
	52	43	44
	52	58	40
	50	50	52
	60	56	42
	54	56	50
	46	40	49
	48	48	44
	48	48	48
	60	54	42
	52	49	52
	59	58	47
	59	53	48
	60	56	44
	60	53	49
Mean value	54,0666667	52	46,4666667

East side of Griegaksla:

	By the beach, 10 m a.s.l.	**By the end of plateau II, 20 m a.s.l.**	**In the middle of the slope, 40 m a.s.l.**
	57	56	60
	45	52	50
	58	54	56
	48	50	47
	58	53	55
	54	62	61
	52	56	54
	58	58	54
	51	54	50
	47	56	52
	50	63	51
	46	55	54
	54	54	50
	53	48	52
	54	59	59
			54
			47
			60
			49
			54
Mean value	52,3333333	55,3333333	53,45

Appendix 4: Established fixpoints, Svalbard

Coordinates for established fixpoints by Griegaksla, Svalbard

Fixpoint	**Co-ordinates (In 33 X)**	**Length 1 (m)**	**Length 2 (m)**	**Length 3 (m)**
1	0470415E 8663652N	3,03	4,43	5,73
2	0470397E 8663663N	2,08	2,40	3,17
3	0470385E 8663676N	2,20	3,14	3,64
4	0470385E 8663727N	3,20	4,27	4,98
5	0470380E 8663729N	1,20	1,86	2,06
6	0470390E 8663751N	2,05	2,49	3,35
7	0470404E 8663762N	1,69	2,57	3,71
8	0470415E 8663779N	2,53	2,73	3,79
9	0470425E 8663791N	2,19	3,13	3,94
10	0470432E 8663796N	1,34	2,30	2,65